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REFERENCE

Lane, B. H., 1935, Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C. 35c.)

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MAY 1956

NEW RADIOCARBON DATES AND LATE-PLEISTOCENE STRATIGRAPHY

RICHARD FOSTER FLINT

ABSTRACT. Carbon 14 dating has become a valuable means of checking, and in some cases correcting, stratigraphic correlations based on field evidence. A series of dates obtained in the Yale Geochronometric Laboratory, in conjunction with dates of related samples published by other laboratories, contribute to a better understanding of late-Pleistocene stratigraphy of northern North America. Specifically the results fix the time of the Valders glacial maximum at around 10,700 yr B.P., improve the dating of phases of the glacial Great Lakes, advance knowledge of glacial events in central Connecticut, indicate at least local glaciation, northwest of Hudson Bay, subsequent to the Thermal maximum, and add to our information concerning a widespread glaciation that affected North America at a time more than 30,000 years ago.

INTRODUCTION

In 1951 Flint and Deevey (1951) discussed the stratigraphic implications of a group of samples, radiocarbon-dated by W. F. Libby, bearing on late-Pleistocene events. The present discussion is similar. It is based on samples dated in the Yale Geochronometric Laboratory. All but three of the dates were published by Preston, Person, and Deevey (1955) under the designation "North American Geology." Originally this group of samples was much larger. While the Geochronometric Laboratory was changing over from solid-carbon counting to gas counting, a number of samples expected to antedate Mankato time were sent to the U. S. Geological Survey Laboratory in Washington, where they were grouped with other "old" samples and dated. Their dates were discussed by Flint and Rubin (1955) and by Flint (1955).

The group reviewed in the present paper is therefore residual, including "young" samples and those "old" samples not sent to Washington. Added to it are two pertinent samples dated in the Chicago laboratory and not previously discussed, otherwise than briefly in published laboratory date lists and by Flint (1953); added also are pertinent dates from the Michigan laboratory. Like its predecessors this report is only a report of progress. It continues a policy of making available significant groups of new dates, accompanied by pertinent references and more discussion than can be given in laboratory lists of dates.

The discussion is predicated on the assumption that the dates dealt with are correct, except those for which reason for doubt is expressed. However, possible errors (see Flint and Rubin 1955, p. 649) may invalidate any date, and we must face the likelihood that one or more of the dates treated herein are seriously wrong, and that inferences drawn from them now must be discarded later.

CHECK LIST OF SAMPLES

(Arranged in order of laboratory number. The errors quoted are derived as explained in Preston, Person and Deevey (1955). Dates obtained by the solid-carbon method are inclosed in brackets; those obtained by the acetylene method are not inclosed.)

Number	Description	Date (C^{14} yr before A.D. 1955)
Y-11	Wood from alluvium of Cypress R., Manitoba	2830 ± 130
Y-19	Driftwood from emerged marine sediment, Brönlunds Fjord, N.E. Greenland	5870 ± 100
Y-147X	Log from Mankato till, Green Bay, Wisconsin	[11,940 ± 390]
Y-165	Peat underlying alluvium related to L. Agassiz 2, Rossendale, Manitoba	12,400 ± 420 [13,230 ± 600]
Y-166	Shells of freshwater clams from alluvium related to L. Agassiz 2, Rossendale, Manitoba	[11,230 ± 480]
Y-169B	Wood from center of peat bed, South Haven, Michigan. Same as M-290 and C-848.	5660 ± 100
Y-215	Shells from Champlain Sea sediments, Hull, Ontario	10,630 ± 330 [11,050 ± 400]
Y-216	Shells from Champlain Sea sediments, Ottawa, Ontario	10,850 ± 330 [8960 ± 380]
Y-222	Peat, overlying clay, Dugwal, Ontario	6730 ± 200 [6970 ± 310]
Y-227	Wood from type locality of Two Creeks peat layer, Two Creeks, Wisconsin	11,130 ± 350
Y-231	Peat underlying till, Rankin Inlet, Northwest Territories	5220 ± 340
Y-233	Shells from Champlain Sea sediments, Montreal, Quebec	11,370 ± 360
Y-237	Log from lacustrine sediment, Menasha, Wisconsin	11,690 ± 370
Y-238	Driftwood from shallow trench, Marinette, Wisconsin	4880 ± 190
Y-242	Wood underlying varved silt, St. Pierre-les-Becquets, Quebec	>30,840 >29,630
Y-250	Wood overlying outwash and overlain by till, Hibbing, Minnesota	>36,490
Y-251	Log from alluvium overlying varved clay and silt, Hartford, Connecticut	10,650 ± 320
Y-253	Wood fragments from gravel, New Britain, Connecticut	10,710 ± 330
Y-254	Wood overlain by till and stratified sediments, Les Vielles Forges, Quebec	>29,630
Y-255	Peat from horizon of Y-254	>30,840
Y-256	Peat and wood overlain by silt and sand, Pierreville, Quebec	>29,630
Y-261	Peat overlain by lacustrine silt and outwash, middle Back River region, District of Keewatin, Northwest Territories	4140 ± 150
Y-269	Peat overlain by till and stratified sediments, Missinaibi River, Ontario	>29,630
Y-270	Wood underlying horizon of Y-269, and separated from it by a thin layer of till	>30,840
Y-271	Shells from horizon overlying marine clay and overlain by sand, Breakneck Falls, Opasatika R., Ontario	17,000 ± 370

Y-282	Gyttja from upper part of Durham spruce-pollen zone, Durham, Connecticut	8155 ± 410
Y-283	Driftwood from Beaufort fm., Intrepid Inlet, Prince Patrick Island, Northwest Territories	>25,300
Y-284	Driftwood from Beaufort fm., Landing Lake, Prince Patrick Island, Northwest Territories	>31,840
Y-285	Gyttja from lower part of Durham spruce-pollen zone, Totoket bog, North Branford, Connecticut	13,550 ± 460
Y-293A	Peat from basal part of peat layer, South Haven, Michigan. Same as M-288.	10,790 ± 200
Y-293B	Wood from surface of unconformity between Glenwood and Calumet phases of L. Chicago. South Haven, Mich. Same as M-288*a* (11,200 ± 600) and W-167 (10,860 ± 350).	10,550 ± 150

TWO CREEKS HORIZON

The Two Creeks horizon, exposed in eastern Wisconsin (fig. 1), and marking the short interval of greatest deglaciation between the Cary and Valders[1] glacial advances (Flint and Deevey, 1951, and references cited therein), was dated by Arnold and Libby (1951, p. 117) at 11,404 ± 350 yr B.P., a mean of 5 individual dates. The mean of two Washington dates (W-42, W-83; Suess, 1954, p. 471) is 11,370 ± 100. A sample of spruce wood (Y 227), collected by F. T. Thwaites from this horizon from the same (the type) locality was dated at 11,130 ± 350 yr, in fairly satisfactory agreement with the values arrived at by the other laboratories.

As shown by Iversen (1953; see also Suess, 1954, p. 472) the Two Creeks dates agree with those of the Alleröd horizon in Europe, which appears

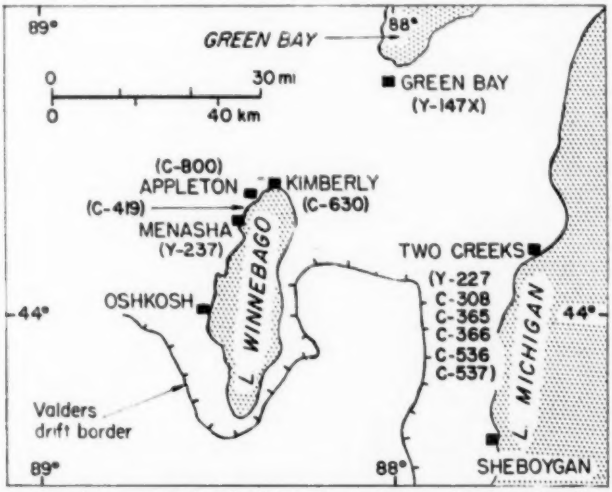


Fig. 1. Sketch map showing localities of occurrence of radiocarbon-dated samples in eastern Wisconsin.

[1] The name Valders (Thwaites, 1943) is used in preference to Mankato for reasons explained in the present paper.

to represent a span of at least 1000 yr. The dates of two samples from other places in Wisconsin (fig. 1) that had been expected to correlate with the Two Creeks horizon confirm expectation in that they agree reasonably well with dates from the type locality of that horizon. Both were submitted by F. T. Thwaites. One, Y-147X, is a transported log that was imbedded in red Valders till at Green Bay, Wisconsin, 33 mi. northwest of Two Creeks; its date is 11,940 ± 390. The other, Y-237, is likewise a transported log inclosed in the same red till at Menasha, Wisconsin, 29 mi. southwest of Green Bay and 44 mi. west of Two Creeks, and collected in 1952 from an industrial excavation then in progress. The date of the Menasha sample is 11,690 ± 370. The log was of oval cross section, inferred by Mr. Thwaites to be the result of deformation antedating the time of its collection.

The Menasha sample has additional interest in that the locality of its occurrence is only 2 mi. south of that of sample C-419 (Arnold and Libby, 1951, p. 118). As discussed by Flint and Deevey (1951, p. 281) Mr. Thwaites had predicted a Two Creeks age for it, but its date was determined as 6401 ± 230. The result was incompatible with the probabilities. On the one hand the log was white spruce and possessed thin growth rings, these two facts implying a climate colder than the present climate of the district. Its cell structure had been crushed, and it was believed to have been inclosed in the till,[2] formed when the Valders glacier, at its maximum, blocked the Green Bay depression. On the other hand the date was some 4000 years younger than the estimated date of the Valders maximum in the district, and hence implied a climate of distinctly nonglacial character. The date of Y-237 suggests that the discrepancy is probably the result of contamination or laboratory error in the dating of C-419, a hypothesis that had been recognized by Flint and Deevey.

The likelihood that this is the case is strengthened by the dates of two other samples from this same district. One, C-630 (Libby, 1952, p. 678) was a piece of driftwood(?) inclosed in varved sediments of glacial lake Oshkosh at Kimberly, Wisconsin, 8 mi. northeast of Menasha. The date, 10,676 ± 750 yr, is compatible with the hypothesis that the wood was growing in the vicinity at about the time when the Valders readvance reached its maximum extent. It is not implied that the wood samples correlated with the Two Creeks horizon were from trees that grew at a single locality, nor that all are of exactly the same age. It is expectable that, following retreat from the Port Huron moraine (pl. 1), much of northeastern Wisconsin would have been covered with subarctic forest, which would have been progressively invaded and destroyed by the Valders glacier. Although the time limits are not precisely fixed, the C^{14} dates suggest that in this part of Wisconsin the retreat from the Port Huron position and the readvance to the Valders maximum might have embraced 1000 to 1500 years.

The other supporting sample is C-800 (Libby, 1954a, p. 139), a spruce log, still retaining some bark, imbedded in red Valders till overlain by sedi-

[2] Not the sediments of glacial Lake Oshkosh, as stated by Flint and Deevey, (1951, p. 281) on information from Mr. Thwaites. Notes made when sample C-419 was collected were lost, and were recovered after 1951; they state that the sample came from till (F. T. Thwaites, written communication, Aug. 15, 1955).

ments of glacial lake Oshkosh at Appleton, Wisconsin, 5 mi. north of Menasha. The sample was collected in 1952 by W. F. Read, generically identified by L. R. Wilson, and submitted for dating by Leland Horberg. Two dating runs on this sample gave 11,471 ± 500 yr and 10,241 ± 650 yr, respectively, with an average of 10,856 ± 410 yr. As Professor Read inferred, the presence of bark implied that the log had not been transported far, and therefore that the tree probably had grown in the vicinity.

In summary, the known samples from eastern Wisconsin give a consistent reconstruction of events during the Cary-Valders interval, the advance to the Valders glacial maximum, and the beginning of the subsequent retreat, provided the date of C-419 is neglected.

Dating of the Valders maximum in eastern Wisconsin, most closely approximated by sample C-630 at around 10,700 yr, compels a second look at the stratigraphic placing of the outwash sediment at Bonfils, Missouri. A driftwood log (C-385) was dated at 12,148 ± 700 yr (Arnold and Libby, 1951, p. 118). The inclosing outwash had been assigned by Leighton and Willman to the Mankato substage, a correlation accepted tentatively by L. C. Peltier, collector of the sample, and by Flint and Deevey (1951, p. 267-268), at a time when few reliable dates older than those of the Two Creeks material had become available.

Now, however, the dates of pre-Two Creeks samples, for example W-33, W-140, and W-161 (table 1), imply that the Bonfils log antedates the Valders maximum by about 1400 yr. If the log was in its primary sedimentary position when found, its date implies that the outwash inclosing it was deposited during shrinkage of the ice sheet toward its Two Creeks minimum. Probably the outwash body was a single stratigraphic unit deposited continuously by meltwater before, during, and after the Wisconsin maximum at 18,000 to 19,000 yr B.P. inferred by Flint and Rubin (1955).

LAKE CHICAGO, CALUMET PHASE

The history of the successive phases of the ancient Great Lakes has been made more clear, and has been corrected at various points, by important further field study (Hough, 1953, 1955) and by radiocarbon dating (e.g., Flint and Deevey, 1951, p. 283-285; Flint and Rubin, 1955; Flint, 1953, p. 912; 1955; Zumberge and Potzger, 1955). Those pertinent dates which are known to me are assembled in table 1, in order of increasing age.

The peat locality at South Haven, Michigan (Zumberge and Potzger, 1955) has yielded a sample that apparently dates the Calumet phase of Lake Chicago. The sample is a piece of wood from the interface of unconformity between silt attributed to the Glenwood phase and overlying sand and pebbles ascribed to the Calumet phase, and occurred at approximately 580 ft. altitude. The interface exposed at the peat locality represents the Bowmanville low-water phase, during which the Two Creeks peat bed accumulated elsewhere. Hence the wood sample could represent any part of the related interval. Zumberge and Potzger (1955, p. 310) placed it simply as Bowmanville. The sample was divided and was dated by the Michigan, Washington, and Yale laboratories. The results (table 1) are in good agreement, with a weighted average value of 10,640 ± 130 yr. The date suggests that the driftwood marks precisely

TABLE 1

Radiocarbon Samples Pertinent to the History of the Great Lakes
(Listed in order of increasing age)

Laboratory number*	Description	Date, C^{14} yr.**	Present stratigraphic assignment (lake phase)	References
C-608	Charcoal, Burley site, Port Franks, Ontario	[2619 ± 220]	Algoma	Libby, 1952, p. 674; Dreimanis, 1952.
C-364	Driftwood, lake sediments, Dolton, Illinois	[3469 ± 230]	Nipissing	Arnold and Libby, 1951, p. 115; Flint and Deevey, 1951, p. 284; Flint, 1953, p. 917.
C-504	Peat, Sand Island, Bayfield Co., Wisconsin	[3656 ± 640]	Nipissing	Arnold and Libby, 1951, p. 118; Flint and Deevey, 1951, p. 284; Flint, 1953, p. 917.
M-291	Peat (uppermost 2 in.), South Haven, Michigan	4000 ± 350	Chippewa-Nipissing transition	Zumberge and Potzger, 1955.
C-849	Idem	[4816 ± 290]		Libby, 1954b, p. 735.
Y-238	Driftwood, lake sediments, Marinette, Wisconsin	4880 ± 190	Nipissing	Preston, Person and Deevey, 1955, p. 958.
C-191	Charcoal, Frontenac Island site, Cayuga Lake, New York	[4930 ± 260]	Post-Iroquois; no upper limit implied	Arnold and Libby, 1951, p. 114; Flint and Deevey, 1951, p. 283.
Y-169B	Wood from peat (central part), South Haven, Michigan	5660 ± 100	Chippewa	Preston, Person, and Deevey, 1955, p. 958.
M-290	Idem	5090 ± 300		Zumberge and Potzger, 1955.
C-848	Idem	[6440 ± 230]		Libby, 1954b, p. 735.
M-289	Peat (7 in. above base), South Haven, Michigan	6330 ± 400	Post-Algonquin, pre-Chippewa	Zumberge and Potzger, 1955.
Y-293A	Peat (lowest 2 in.), South Haven, Michigan	10,790 ± 200	Algonquin (late)	Preston, Person, and Deevey, 1955, p. 958 (therein given incorrectly as 9500 ± 250).
M-288	Idem	7925 ± 400		Zumberge and Potzger, 1955.
C-846	Idem	[6744 ± 530]		Libby, 1954b, p. 735.
C-674	Driftwood, lake sediments, Chicago, Illinois	[8200 ± 480]	Tolleston	Libby, 1952, p. 674; Flint, 1953, p. 917.
C-526	Wood, forest horizon overlain by lake sediments, Castalia, Ohio	[8513 ± 500]	Lundy	Libby, 1951, p. 292; Flint, 1953, p. 917.

TABLE 1 (Continued)

Laboratory number*	Description	Date C^{14} yr.**	Present stratigraphic assignment (lake phase)	References
W-199	Driftwood, lake sediments, Marilla, New York	9640 ± 250	Warren	Rubin and Suess, 1955, p. 483.
Y-293B	Driftwood, lake sediments, South Haven, Michigan	10,550 ± 150	Calumet	Preston, Person, and Deevey, 1955, p. 958.
W-167	Idem	10,860 ± 350		Rubin and Suess, 1955, p. 483.
M-288*a*	Idem	11,200 ± 600		Zumberge and Potzger, 1955.
C-308, 365 366, 536, 537	Wood and peat, Two Creeks horizon, Two Creeks, Wisconsin	[11,000 to 12,000]	Bowmanville	Arnold and Libby, 1951, p. 117.
W-42, 83; Y-227	Idem	11,000 to 11,500		Suess, 1954, p. 471; Preston, Person and Deevey, 1955, p. 958.
W-161†	Driftwood, within second spit, Dyer, Indiana	12,200 ± 350	Glenwood	Rubin and Suess, 1955, p. 483; Flint and Rubin, 1955, p. 650.
W-140‡	Driftwood, lagoon behind first spit, Dyer, Indiana	12,650 ± 350	Glenwood	Idem; idem.
W-33	Driftwood trash, lagoon sediments, Cleveland, Ohio	13,600 ± 500	Arkona	Suess, 1954, p. 469; Flint and Rubin, 1955, p. 649-650

* C = Chicago; M = Michigan; W = Washington; Y = Yale.

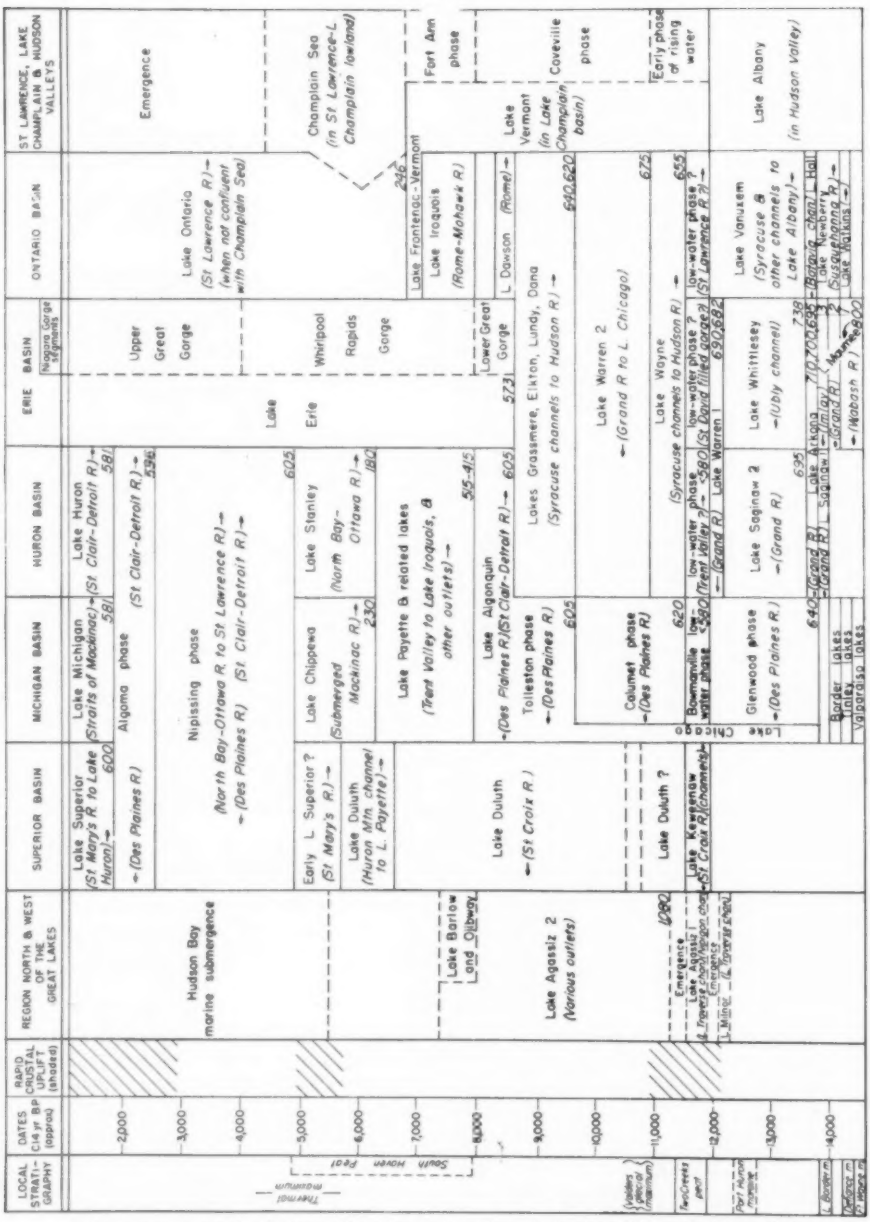
** Solid-carbon dates are inclosed in square brackets; all others are based on gas counting.

† A sample (C-801; Libby, 1954a, p. 137) taken earlier from the same stratigraphic position as W-161, and at the same locality, dated 10,661 ± 460 and 11,284 ± 600, averaging 10,972 ± 350. As the dates were considered to be too young to match the stratigraphic position, sample W-161 was collected; its date is accepted here.

‡ Two samples (C-871, C-872; Libby, 1954b, p. 735) taken earlier from the same stratigraphic position as W-140, and at the same locality, dated 18,500 ± 500 and >21,000 respectively. As the dates were considered to be too old to match the stratigraphic position, sample W-140 was collected; its date is accepted here.

the close of the Bowmanville phase at the locality, having been tossed up and incorporated into the beach built by the rising water of the Calumet phase 40 ft below the maximum height of that phase. If such is the case the wood is better described as early Calumet. Its date falls nicely between the group of Two Creeks dates and the date of a Lake Warren driftwood sample (9640 ± 250; table 1 and pl. 1). Also it approximates that of the Valders maximum. This suggests that although the Valders glacial advance had closed the Straits

PLATE 1


LOCAL STRATIGRAPHY
DATES C14 yr BP (approx)
RAPID CRUSTAL UPLIFT (shaded)
REGION NORTH & WEST OF THE GREAT LAKES
SUPERIOR BASIN
MICHIGAN BASIN
HURON BASIN
ERIE BASIN
ONTARIO BASIN
ST LAWRENCE, LAKE CHAMPLAIN & HUDSON VALLEYS
Thermal maximum
South Haven Peat
Two Creeks peat
Port Huron moraine
Mackinac moraine
Valparaiso moraine
Hudson Bay marine submergence
Lake Barlow Land Olbway
Lake Agassiz 2 (Various outlets)
Lake Agassiz 1
Emergence
Lake Superior (St Mary's R. to Lake Huron)
Early L Superior? (St Mary's R.)
Lake Duluth (Huron Mtn channel to L. Payette)
Lake Duluth
Lake Duluth ?
Lake Keweenaw
Lake Michigan (Straits of Mackinac)
Algoma phase
Nipissing phase
(North Bay-Ottawa R. to St Lawrence R.)
(Des Plaines R.) (St Clair-Detroit R.)
Lake Chippewa
(Submerged Mackinac R.)
Lake Payette & related lakes
(Trent Valley to Lake Iroquois, & other outlets)
Lake Algonquin
(Des Plaines R./St Clair-Detroit R.)
Toleston phase
Calumet phase
Bowmanville low-water phase
Glenwood phase
Lake Huron
(St Clair-Detroit R.)
Lake Stanley
(North Bay-Ottawa R.)
Lakes Grassmere, Elkton, Lundy, Dana
(Syracuse channels to Hudson R.)
Lake Warren 2
(Grand R to L. Chicago)
Lake Wayne
(Syracuse channels to Hudson R.)
low-water phase ?
(St Clair filled gorge)
Lake Warren 1
Lake Saginaw 2
(Grand R.)
Lake Saginaw 1
Lake Arkona
Lake Whittlesey
(Ubly channel)
Lake Iroquois
(Rome-Mohawk R.)
L Dawson (Rome)
Lake Vanuxem
(Syracuse & other channels to Lake Albany)
Lake Frontenac-Vermont
Champlain Sea
(in St Lawrence-L Champlain lowland)
Fort Ann phase
Lake Vermont (in Lake Champlain basin)
Coveville phase
Early phase of rising water
Lake Albany
(in Hudson Valley)
Emergence
Niagara Gorge segments
Upper Great Gorge
Whirlpool Rapids Gorge
Lower Great Gorge
Lake Ontario (St Lawrence R.) (when not confluent with Champlain Sea)
Lake Erie
Chicago outlet
Border lakes
Tinley lakes
Valparaiso lakes
Grand R.
Saginaw Bay
Mohawk R.
Susquehanna R.
Lake Watkins
Wabash R.
Ohio R.
Mississippi R.
Hall
Rochester
Iroquois R.
Champlain
Hudson R.
Albany
Trenton
Ottawa R.
Des Plaines R.
St Croix R.
St Mary's R.
Straits of Mackinac
Trent Valley
Huron Mtn channel
Keweenaw
Lake Michigan
Lake Huron
Lake Stanley
Lake Chippewa
Lake Superior
Lake Duluth
Lake Algonquin
Lake Warren 2
Lake Wayne
Lake Saginaw 2
Lake Whittlesey
Lake Vanuxem
Lake Albany
Lake Iroquois
Lake Frontenac-Vermont
Champlain Sea
Lake Ontario
Lake Erie
Lake Vermont
Fort Ann phase
Coveville phase
Early phase of rising water
Emergence
Lake Barlow Land Olbway
Lake Agassiz 2
Lake Agassiz 1
Hudson Bay marine submergence
Nipissing phase
Algoma phase
Toleston phase
Calumet phase
Bowmanville low-water phase
Glenwood phase
Lakes Grassmere, Elkton, Lundy, Dana
Lake Warren 2
Lake Wayne
Lake Saginaw 2
Lake Whittlesey
Lake Vanuxem
Lake Albany
Lake Iroquois
Champlain Sea
Lake Ontario
Lake Erie
Lake Vermont
Fort Ann phase
Coveville phase
Early phase of rising water
Emergence

of Mackinac prior to 11,400 yr ago, thereby causing the surface of the lake to rise from its Bowmanville position (at an unknown altitude of less than 580 ft), the advance approached its maximum near Milwaukee, several hundred years later, before the water in the Lake Michigan basin rose to the overflow point, the Des Plaines River spillway at 620 ft. There is an implication here that the Bowmanville water level may have been much, rather than only a little, below 580 ft.

NIPISSING GREAT LAKES

Although the general agreement of many of the various Great Lakes dates is satisfactory, there is an unresolved discrepancy in the Algonquin-Chippewa-Nipissing segment of the table. This can be seen on plate 1, a revision of an earlier-published chart (Flint, 1953, pl. 3), improved mainly through the incorporation of data from a valuable paper by Hough (1953) and through the adoption of a time-calibrated vertical scale.

In the two instances in which replicate dates on the same sample are in conflict, the chart incorporates a choice based on weight of evidence. The choices can be only tentative, and are subject to modification when future data permit wider choices. As a post-Algonquin date, 7925 yr (M-288) is preferred to 6744 (C-846) because it better fits[3] the close of the spruce-fir forest, with which it is pollen-correlated, and to the date of Y-293A, obviously an erroneous value; although it allows a surprisingly short time for Lake Algonquin, which can not have begun until after 9640 yr B.P. (W-199).

Also 4800 yr (C-849) is preferred, as a date near the beginning of the Nipissing phase, to 4000 yr (M-291) because the date of Y-238, a probable Nipissing event, is in good agreement. We note also that the 4800-yr date is not in conflict with the correlation of the Nipissing with the Thermal maximum (Xerothermic) (Zumberge and Potzger, 1955, p. 310).

Sample Y-238 is pertinent to this problem. It is a piece of driftwood (white oak; Thwaites, 1943, p. 141) collected prior to 1943 by the City Engineer of Marinette, Wisconsin, from a sewer trench and retained for many years in the Department of Geology of the University of Wisconsin. According to F. T. Thwaites, who submitted the sample, the altitude of occurrence must lie between 580 ft and 600 ft, and the wood probably was emplaced during the Nipissing phase, either in the lake, or in the mouth of the Menominee River when the lake receded at the close of Nipissing time. Its date, 4800 $\pm$ 190, is compatible with the first alternative, and agrees with the Nipissing date (C-849) determined by Libby (1954b, p. 735). Although it falls within the Chippewa phase in the scheme of Zumberge and Potzger, it is tentatively accepted as an early Nipissing date for the reasons stated. Furthermore it affords a more reasonable minimum lower limit (around 5000 yr B.P.) for the Thermal maximum than does the Zumberge-Potzger scheme, in that according to those authors pollen evidence shows the Thermal maximum to have been coincident with the Nipissing phase.

The discrepancy remains unresolved and very likely will retain that status until more critical data are published.

[3] In the opinion of Professor E. S. Deevey (unpublished).

LAKE AGASSIZ

Events in the history of Lake Agassiz are less well dated than those pertaining to the glacial Great Lakes farther east; hence sample Y-165 holds unusual interest and poses a problem. It is from a layer of peat, overlain by 13 ft of alluvium identified by J. A. Elson (unpublished) as part of a delta built into Lake Agassiz 2. This was the later phase of the lake created, after an interval of desiccation, by a glacial readvance tentatively correlated by Elson with the Valders advance in the Lake Michigan basin. The date of the sample, 12,400 ± 420 yr, is compatible with the correlation suggested. Another sample, Y-166, consists of shells of freshwater clams from the alluvium that overlies the peat, exposed without the peat at a second locality nearby. This sample was dated, by the solid-carbon method, at 11,230 ± 480. The two dates are internally consistent.

They are consistent also with C-497 (Arnold and Libby, 1951, p. 117; Rosendahl, 1948, p. 289-290; Flint and Deevey, 1951, p. 282), wood that apparently underlay 25 ft of Lake Agassiz silt at Moorhead, Minnesota, and that dated at 11,283 ± 700 yr. Although there is doubt as to the exact stratigraphic position of the wood, reported depth below the surface and lithology of the inclosing sediment would place the wood in unit 4 of Rominger and Rutledge (1952, table 3, p. 168) at the adjacent locality of Fargo, North Dakota; this unit, in turn, is correlated by Elson with Lake Agassiz 2.

All three samples are in apparent conflict with a group of five dated samples from two localities in central Iowa (C-596, C-653, C-664, C-912, C-913; Libby 1952, p. 675; 1954b, p. 737) described by Ruhe and Scholtes (1955). Three of the samples were found in or beneath till classified as Cary; the other two (hemlock wood recording a climate cooler than that now existing) were found in drift classified as Cary or possibly pro-Mankato. The dates, ranging from about 12,000 to 14,000 C^{14} yr appear to be reasonable Cary dates, as Ruhe and Scholtes remarked.

However, the older of the two Lake Agassiz dates, that of Y-165, falls within the range of the five Iowa dates. The sediments inclosing and overlying the sample postdate the latest glaciation of the district; yet sediments of apparently comparable age in Iowa imply the presence of a glacier 600 mi. farther south. The discrepancy may result merely from probable counting errors, with the consequence that the implied overlap is not real but only apparent. Furthermore the 600-mi. retreat from Iowa to central Manitoba may have been rapid, comparable with implied rates in the Ohio-Michigan region during the contemporaneous rapid retreat from the Port Huron moraine to the Straits of Mackinac (Flint, 1955, fig. 2). Clarification awaits further stratigraphic study and additional dated samples.

A minimum terminal date for Lake Agassiz is implied by sample Y-11, wood from alluvium of Cypress River, Manitoba, evidently postdating the lake. The date is 2830 ± 130 yr, and indicates a climate more moist than that of the present time. Samples from a comparable stratigraphic position, alluvium unconformably overlying Lake Agassiz sediments at a locality in northern North Dakota (C-722 : 2150 ± 400 yr; C-723 : 2684 ± 200 yr, unpublished) gave dates in agreement with that of Y-11.

RELATIONS OF TYPE MANKATO, VALDERS, AND PORT HURON DRIFT SHEETS

If the Iowa and Lake Agassiz dates discussed in the preceding section are even approximately correct, the implication appears that the youngest drift of the Des Moines lobe in Iowa is older than the Valders drift overlying the Two Creeks horizon in eastern Wisconsin. This possibility, mentioned by Horberg (1955), raises a question as to the date and correlation of the Mankato drift sheet at its type locality in southern Minnesota. Not dated as yet at that locality, the Mankato drift has long been considered equivalent to the youngest drift in north-central Iowa. If equivalence should be confirmed, the Valders drift sheet would represent, not the (type) Mankato maximum (see Wright, 1955), but a glacial advance younger than that maximum by perhaps as much as 2000 yr. A further probable corollary of the correlation is that the Port Huron moraine and drift sheet, dated at some time between 13,600 and 12,200 years B.P. (Flint and Rubin, 1955, p. 650), is the approximate equivalent of the Mankato maximum (although not of the Valders), rather than being a "late Cary" feature as interpreted by Bretz (1951, p. 424; see also p. 412). Because of these probabilities the term "Mankato" has been omitted from the chart (pl. 1).

CONNECTICUT

It was tentatively suggested (Flint, 1953) that a glacial readvance reaching to Middletown, Connecticut, represents the Cary[4] maximum. On certain assumptions the readvance was dated at 20,000 yr B.P., on the evidence of the radium date of a varve lamina at Hartford and two radiocarbon dates of Cary(?) drift in Ohio. Subsequently abundant additional evidence from Ohio (see Flint and Rubin, 1955) indicated that the drift in that State, represented by the two samples dated earlier, is not Cary(?) but pre-Cary. Furthermore, radiocarbon dates of Connecticut samples from pollen-studied sequences, discussed below, are in conflict with the radium date of the varve lamina. They imply that the glacial readvance to Middletown occurred before, but probably not long before, about 13,000 yr ago, a time in reasonable agreement with Cary events in the Great Lakes region (Flint and Rubin, 1955, p. 649-650). Hence the newer information reduces the absolute date of the readvance, but supports its correlation with the Cary drift. As the Cary maximum has not yet been accurately radiocarbon-dated in the Great Lakes region, it is not known whether the Middletown readvance represents the Cary maximum or a later event.

Four samples from pollen-studied sequences in Connecticut date events in a well established climatic history. The most recent contribution to this history is the work of Estella Leopold, and forms the subject of a paper soon to be published. The stratigraphy of an extinct lake at Durham (4 mi.) and of a bog at Totoket (17 mi.) south of the limit of the Middletown readvance records the high-pine pollen Zone B of Deevey (1939, 1943). The Durham section dates that zone at 8155 ± 410 yr (Y-282), in excellent agreement with its date at Upper Linsley Pond (8323 ± 400) (C-39; Arnold and Libby,

[4] The term "Cary" refers to the drift at the type locality at Cary, Illinois. It is used here only to facilitate discussion of the 1953 paper referred to.

1951, p. 114; Flint and Deevey, 1951, p. 271), close to the Totoket locality. The sequence also records, immediately below Zone B, two distinct spruce maxima, the lower of which dates 12,700 ± 280 (W-46: Suess, 1954, p. 468) at Durham and 13,550 ± 460 (Y-285) at Totoket, a good agreement. At both localities the lower one of these two spruce maxima is underlain by a zone rich in non-arboreal pollen and recording climatic chilling. Dr. Leopold correlated the chilling with the Middletown readvance, apparently a Cary event, and the upper spruce maximum, assumed to be equivalent to Maine Zone L-3 of Deevey (1951), with the Valders maximum.

Y-253, collected by Kurt Servos, Estella Leopold, and R. F. Flint, consists of broken segments of small twigs occurring in gravel 3000 ft south of the mouth of the New Britain channel (Flint, 1933, p. 975) in the northern part of the town of Berlin. The gravel body, 9 ft thick, unconformably overlies the Berlin clay (Flint, 1933, p. 968) and is overlain by 1 to 2 ft of clayey silt, apparently water-laid, which constitutes the present surface. The gravel consists mainly of pebbles, diameter about 1/4 in., predominantly of the local bedrock. It possesses cut-and-fill stratification with courses 1 to 2 ft thick, and is foreset toward the southeast. The bits of twigs, carbonized and apparently burned, were lying parallel with the foreset laminae, obviously emplaced with the surrounding pebbles. They were seen only in a thin zone about 4 ft below the top of the gravel body, which is estimated to lie about 5 ft lower than the floor of the New Britain channel. The wood was submitted to Dr. W. L. Stern of the Yale School of Forestry for taxonomic identification, which proved to be impossible because of the carbonized condition of the twigs. Consequently no climatic inference can be drawn from the sample itself.

In the opinion of Dr. R. E. Deane, who is currently mapping the Pleistocene geology of the district, the gravel was deposited essentially as a fan by the stream spilling out of the Hartford lake through the New Britain channel. The presence of the lake, constituting a settling basin at the channel head, and the local lithology of the pebbles suggest that the gravel was derived from erosion of the channel during outflow. The thickness of the lake deposits, locally exceeding 250 ft, implies long occupation of the outlet, but the time relations of the gravel within the span of that occupation are not known.

It seems hardly possible that the gravel could postdate the outflow, as it is very improbable that the channel has been occupied by a through stream since the Hartford lake ceased to exist.

The date of the sample is 10,710 ± 330 yr. If the inclosing gravel does in fact record a time during the existence of the Hartford lake, it is compatible, as far as it goes, with W-46, the upper spruce maximum at Durham, which records a time during retreat of the ice-sheet margin following its readvance to Middletown. The Hartford lake formed early in that retreat and, as inferred from the more than 300 ft of varved silt and clay deposited in it, endured throughout a considerable period. The date of Y-253 approximates that of the Valders maximum in eastern Wisconsin (as implied by C-630 and C-800), a time when presumably the ice-sheet margin stood in extreme northern New England (Flint, 1953). However, neither the position of the glacier

margin nor that of the northern end of the lake at the time of drainage is known from sedimentary or morphologic evidence, and it should be emphasized that the lake was destroyed by erosion of the dam at its southern end, rather than by events at its northern end.

The date of destruction of the lake is, however, limited by the date of Y-251, a driftwood log, minus its bark, imbedded in terrace alluvium occurring along the north branch of Park River in the northwestern part of the city of Hartford. The alluvium caps a stream terrace that truncates the varved silt and clay deposited in the Hartford lake. At the time the log was emplaced, the lake had been drained and the Connecticut River and its tributaries were dissecting the varved sediments; about 40 ft of incision had occurred, of a total of at least 100 ft of measured post-lake incision.

The wood, collected by R. E. Deane, was identified by W. L. Stern of the Yale School of Forestry, as *Picea* sp. It represents a tree that must have grown within the drainage basin of the local stream, within a few miles of the locality at which it was found (unless, of course, the log at its place of occurrence was redeposited, a possibility that can not be excluded). Ecology and climatic implications can hardly be inferred from a single occurrence; however, the presence of spruce in central Connecticut at the date of the sample, 10,650 ± 320 yr B.P., is compatible with Dr. Leopold's pollen sequences.

The dates of these two samples imply that the episode of draining of the Hartford lake antedated Y-251 and either postdated or accompanied Y-253. If the stratigraphic relations of each have been interpreted correctly, Y-253 should be the older by an amount determined by the particular part of the outflow period during which it was deposited, plus the time involved in dissection of the lake deposits to the point at which the terrace of Y-251 was made.

Calculation from the counting errors indicates a probability of only 1 in 40 that Y-253 is older than Y-251 by as much as 1000 years. The geologic relations, on the other hand, suggest a difference of this order. This is as far as we can go without either additional dates or more refined geologic information.

The date of Y-253 is younger by about 7000 years than the date of varve no. 3700 in the Hartford lake sequence determined, on explicit assumptions, by Urry (1948) from the radium content of the varve. Varve 3700 lies in the upper part of the sequence at Hartford, as can be inferred from the sections published by Antevs (1922). Urry deduced that the site of Hartford was deglaciated about 19,500 years ago.

Obviously the C^{14} dates discussed here do not agree with the radium determination. Because the internal consistency of the C^{14} dates is good, they are accepted tentatively, but with the hope that further radium dates will become available, so that a better comparison can be made.

In summary, the group of Connecticut dates places the glacial readvance to Middletown at some time before 12,700 yr ago; the Hartford lake, formed after the readvance, was still in existence 10,710 yr ago, but had been drained

prior to 10,650 yr ago. The last two dates are uncomfortably close to each other; where the error, if any, lies can not be determined at present.

CHAMPLAIN SEA

The relation of the Champlain Sea to the sequence of glacial Great Lakes (pl. 1) is a matter of much interest. The marine invasion has long been inferred to have postdated Lake Iroquois, but lack of evidence has made closer placement difficult. A concerted attempt was made in 1953 to obtain samples that would permit C^{14} dating of the Champlain marine deposits.

Three samples were obtained and dated. Nos. 215 and 216, collected by J. A. Elson, represent an early phase of the marine invasion in the Ottawa district, Ontario. No. 233, collected by V. K. Prest, and likewise early, is from the Montreal district. Despite a search for wood, the only datable material found consisted of shells; hence all three samples are shells, mostly *Saxicava* and *Macoma*, collected from primary positions in the inclosing fine sediments.

The (acetylene) dates, listed at the beginning of this paper, range from 10,630 to 11,370 years. Although internally consistent, they are incompatible with the dated Great Lakes sequence. The Champlain Sea sediments are apparently younger than those of Lake Iroquois, which, in turn, postdate those of Lake Lundy and of Tolleston Lake Chicago, both of which are dated (C526: Libby, 1951, p. 292; C674: Libby, 1952, p. 674) at 8000-8500 years. The stratigraphic probabilities implicit in the chart, plate 1, lead to an expectable date of around 6000 to 7000 years for the inception of the marine phase. Hence the dates of the three Champlain Sea samples are rejected as being probably spurious.

The most likely source of the inferred error is incorporation in the clam shells, during growth, of "old" carbonate contributed to the Champlain arm of the sea by streams draining wide areas of limestones and other carbonate rocks exposed within the St. Lawrence River drainage basin. An exaggeration of the true date by 50 percent or so through this means is not without precedent, for marine shells known to be modern have yielded C^{14} dates as great as 1900 years (Kulp, 1952, p. 412). Therefore perhaps the tying-in of the Champlain Sea to the Great Lakes sequence can not be confirmed by C^{14} evidence until wood samples have been found and dated. There seems to be no reason why driftwood should not be present in the marine deposits, as it has been collected from several Great Lakes deposits of greater age.

The occurrences of marine deposits underlying till in the St. Lawrence region, reported by Coleman (1927; 1932) and briefly discussed by Flint (1953, p. 906) have not been confirmed by recent field studies in that region (Paul MacClintock and N. R. Gadd, unpublished communications). In other words there is no evidence of marine invasion of the St. Lawrence lowland immediately preceding the Port Huron or the Valders readvance. Unless such evidence is found, we can only suppose that glacier ice remained in the lowland throughout the Cary-Valders interval, and did not permit a marine invasion until Champlain time, well after the Valders maximum.

The problem of the Ottawa Sea of Antevs and Goldthwait (Antevs, 1939, p. 712), briefly discussed by Flint (1953, p. 910), likewise remains unre-

solved. The exposures on which some of the original observations were made had been destroyed before 1955, when an attempt was made to restudy them. Current field investigations in the region have not revealed further evidence of two marine deposits separated by an unconformity (Paul MacClintock and N. R. Gadd, unpublished communications). In this discussion, therefore, the matter is left in abeyance and the Ottawa Sea is excluded from the chart (pl. 1), without prejudice as to the solution of the problem.

"COCHRANE" PROBLEM

A sample of peat (Y-222) from a bog overlying lacustrine sediments at Dugwal, 34 mi. due south of Cochrane, Ontario, was collected in 1952 by J. A. Elson and H. G. Ignatius in order to establish a minimum date for the latest glaciation of the district south of Cochrane. The date (6730 ± 200 yr) is compatible with that of W-136 (Rubin and Suess, 1955, p. 485) (6,380 ± 350), a sample of peat collected with a similar objective from Cochrane. It implies that the Dugwal locality was last deglaciated more than 6730 years ago, although how much more is not known. Presumably the deglaciation was the one that coincided with Lake Payette and its associates (pl. 1). The date is compatible likewise with that of C-334 (Arnold and Libby, 1951, p. 117), marking pine-pollen Zone B in extreme northeastern Minnesota (Flint and Deevey, 1951, p. 272), 7128 ± 300 yr B.P.

The draining of the lakes of the Payette group and the inception of Lakes Stanley and Chippewa mark the opening, by deglaciation, of the outlet at North Bay, Ontario. This event fixes the position of the glacier margin at North Bay as contemporaneous with the transition from the Payette group to Lake Stanley. North Bay lies 180 mi. south of Cochrane; yet according to the scheme adopted in plate 1, Lake Stanley came into existence around 6000 years B.P. As the date of inception of Lake Stanley is not fixed directly by any sample discussed in the literature, it is possible that the end of the Payette-group phase occurred earlier than is shown on the chart. On the other hand it is entirely possible that the glacier margin retreated rapidly northeastward across the country south of Hudson Bay while ice emanating from the highlands of Quebec, where both altitude and precipitation would favor its persistence, maintained a blockade across the North Bay outlet escape route. If this was the case, there is no discrepancy between Y-222 and M-289.

More importance has been attached to glacial events in the Cochrane district than seems to be warranted by the evidence. As shown by Antevs (1925, p. 75) deglaciation of the wide region of which that district is a part created the large Lake Barlow-Ojibway, which existed for as long as the ice sheet constituted its northern shore. Lacustrine sediments, including varved silt and clay, accumulated in the lake, which was then drained as a result of further deglaciation (Antevs, 1928, p. 103).

According to the literature the ice sheet then re-expanded, depositing till in places on the sediments of the former lake. Till in this position is reported from localities in Ontario by Antevs (1928, p. 104; see also 1931, p. 19) and by Hills (1947). Antevs inferred a minimum readvance of 70 mi. The inferred readvance has been questioned in oral discussions by some geologists, and has

been reaffirmed by others, but the details are not dealt with in the recent literature, and remain unevaluated.

Wide interest in the problem resulted from the publication of a paper (Bryan and Ray, 1940) attempting to reconstruct events since the Wisconsin maximum in eastern North America. This paper, an important milestone in the development of Wisconsin geochronology, has been widely cited. It included two tables (tables 4 and 5) in which the drift representing the readvance in the Cochrane district, based on Antevs' data, was shown as having substage rank, comparable with the Iowan, "Tazewell-Cary," and Mankato substages shown in the same tables, and was correlated with the Fennoscandian "substage" in northern Europe. The "Post-glacial" was shown as beginning with the glacial retreat following the Cochrane readvance. Essentially similar relations were shown in a subsequent publication (Bryan, 1941, p. 57-58), in which the Cochrane "stage" was characterized as "a minor episode of cold" during the period of climatic warming that followed the Mankato maximum.

As far as I have been able to learn, the impression is widespread that there was a significant post-Mankato readvance of the southern margin of the ice sheet, and that the name Cochrane is applied to it. The basis of this impression seems to consist of the Antevs references, plus the two later references, already cited. Whatever the extent and character of the readvance, the propriety of assigning stratigraphic stage or substage value to deposits as little known as those in the Cochrane district may well be questioned. All that the present sample (Y-222) tells us is that glacier ice has not stood over the locality of its occurrence since some time more than 6730 years ago. It is to be hoped that future field study will clarify the late-Wisconsin history of the district.

MISCELLANEOUS SAMPLES OF WISCONSIN AGE FROM NORTHERN NORTH AMERICA

District of Keewatin.—Two samples, both from localities in the tundra region northwest of Hudson Bay in the District of Keewatin, Northwest Territories, are of unusual interest in that they throw light on late Wisconsin events in a region whose glacial history is almost unknown.

One of them, Y-231, comes from a point a few miles north of Rankin Inlet (south of Chesterfield Inlet), at 62°50′ N.Lat.; 92°12′ W.Long. The sample is peat, collected by the late Z. L. Sujkowski from a long trench cut for mineral exploration in perennially frozen ground. The peat forms a layer 6 in. thick, overlain by 4 ft of till capped by patches of thin marine(?) sand. Beneath the peat is chemically decomposed Precambrian sandstone. The peat was protected from erosion by the glacier that deposited the till, by virtue of its situation in the lee of a boss of the sandstone.

The peat clearly antedates the latest glaciation of the locality. The date of the sample, 5220 ± 340 yr, therefore establishes a glacial advance that is late in terms of the Wisconsin age. Clearly, too, the present frozen state of the peat indicates that the temperature of the district was higher than now at the time when the peat accumulated. In view of the date, we can reasonably correlate the peat with the Thermal maximum (see review in Flint and Deevey, 1951, p. 274-281). Glaciation, followed later by freezing of the peat and the

overlying till, probably occurred during the period of lowered temperature that succeeded the Thermal maximum (Flint and Deevey, 1951, p. 276). The beginning of the colder period is estimated at about 4000 years B.P.

The direction of movement of the glacier, reported by Mr. Sujkowski, was toward the southeast; the significance of this observation is discussed below.

The other sample, Y-261, also peat, comes from a bluff exposure on a tributary to the middle segment of Back River at 66°10′ N.Lat.; 97°03′ W.Long., 250 mi. northwest of the locality of Y-231. There the peat forms a thin layer underlain and overlain by lake deposits identified by the collector, R. S. Taylor, as proglacial. Each lake sequence includes a zone of involutions attributed to frost action.

The date is 4140 ± 150 yr. The peat antedates, apparently closely, the latest glaciation of the district (but not of the locality) represented by the overlying lake deposits formed as ice advanced from the southeast. The date is compatible with that of Y-231; the inferred post-peat glaciation could have resulted from the cold period following the Thermal maximum.

According to Mr. Taylor the sediments underlying the peat were deposited in a lake dammed by a glacier that was shrinking toward the southeast. Both above and below the peat, therefore, we have evidence of a source of glacier ice lying southeast of this locality. On the other hand the source of what was apparently the same glacier lay northwest of the Rankin Inlet locality. From these data we can infer a center of glacier outflow lying between the two points perhaps 4000 years ago; the glacier, spreading both ways from the center, covered the Rankin Inlet locality to the southeast but apparently just failed to reach the Back River locality to the northwest.

This inference is doubly interesting in that Y-222 tells us that the Cochrane district, less than 200 mi. south of James Bay, had been deglaciated prior to about 6700 years ago. How much, if any, of the Hudson Bay lowland remained encumbered with glacier ice at the time when a glacier flowed southeast near Rankin Inlet? We do not yet know. However, an ad hoc case can be made out for very rapid deglaciation of the Hudson Bay region during the early part of the Thermal maximum. The Bay region was widely open to the sea at Hudson Strait. When deglaciation of the Strait permitted marine water to leak into the still-depressed Bay region, speedy breakup of the thinned-down ice sheet by calving on a wide front facing northeast should have occurred.[5] The final stages might have witnessed wholesale flotation of the remaining ice. These suggested events would not have affected the ice sheet covering the highlands between Hudson Bay and the Coast of Labrador, other than to cause some calving along its western margin. There seems to be no reason for doubting that the Hudson Bay ice could have been destroyed by this mechanism while the ice farther east remained intact.

Postglacial marine deposits near James Bay.—Another late-Wisconsin event, important and not yet dated, is the marine invasion of the Hudson Bay lowland that accompanied or followed deglaciation. That event has a bearing on the deglaciation of Hudson Bay mentioned in the preceding section.

[5] Originally suggested by A. L. Washburn (unpublished).

A first attempt to date the invasion is represented by sample Y-271, collected in 1954 by Owen Hughes and submitted by V. K. Prest. The locality is near the Breakneck Falls portage on the Opasatika River in northern Ontario, southwest of James Bay at 62 ft altitude. The sample consists of shells, taken from a blanket of marine silt and clay that constitutes the surface material except for a thin deposit of sand.

The date, 17,000 ± 370 yr, is far too old to be acceptable. The composition of the sample, shell rather than wood, makes it suspect for the reason stated in connection with Y-215, 216, and 233. A sample (Y-271) of wood fragments mixed with silt from the same locality, stratigraphic position, and collector as the shells is in the Yale laboratory, but the wood is insufficient in amount for a dating run under present conditions. The unacceptable date of Y-271 should be checked against that of a sample of wood.

Greenland.—Y-19 is driftwood (probably *Larix sibirica*) collected from marine silt and clay in Brönlunds Fjord, Peary Land, north Greenland. According to the collector, J. C. Troelsen (1952, p. 219) the marine sediments mark an interval between two glaciations, the second of which was weak, having been characterized only by a small valley glacier, and was followed by emergence of at least 214 ft.

The date, 5870 ± 100 yr, is close to that of C-358 (Arnold and Libby, 1951, p. 112) (5824 ± 300), marking early Atlantic or late Boreal time in Ireland (Flint and Deevey, 1951, p. 274), and therefore supports a broad correlation with the Thermal maximum. It agrees also with that of Y-231, and like that sample, implies renewed glaciation postdating the sample and therefore the Thermal maximum.

It can be inferred also that at least 214 ft of emergence have occurred within the last 5800 years. Presumably the emergence consists of postglacial crustal recovery minus rise of sea level; until these two factors have been evaluated neither activity can be measured as to rate.

In this respect Y-19 resembles another sample, W-48 (Suess, 1954, p. 468), whale baleen from beach sediments in northwestern Greenland. The sediments lie 44 ft above present sea level and are overlain by 8.5 ft of perennially frozen ground. Both emergence and lowered temperature, therefore, have occurred since the date of the sample, 8500 ± 200 yr B.P. It is implied further that the sea must have been at least moderately clear of floating ice when the whale was alive.

"OLD" SAMPLES

A group of 25 samples is designated here as "old" because the dates of all of them exceed the dates obtained hitherto on materials correlated with the Wisconsin glacial stage (Flint and Rubin, 1955). All the samples in this group were counted by the acetylene method; none showed any activity measurable by the method. This implies that the resulting dates are minimum only; most of the minima fall near 30,000 yr B.P. The differences in the quoted minimum dates of the various samples result from variations in laboratory procedure. In each sample the amount by which the true date exceeds the minimum date is absolutely unknown; a pre-Pleistocene sample would yield a similar result.

TABLE 2
"Old" Radiocarbon Samples

Laboratory number*	Locality (near)	Date (C^{14} yr.)	References**
W-115	Brookings, South Dakota	>30,000	(2), (3)
W-102	Bronson, Minnesota	>36,000	(1), (3), Rosendahl, 1948; Flint and Deevey, 1951, p. 289.
Y-250	Hibbing, Minnesota	>36,490	
W-101	Ironton, Minnesota	>32,000	(1), (3)
W-99	Redwood Falls, Minnesota	>31,000	(1), (3)
W-139	Independence, Iowa	>38,000	(2), (3)
W-67	Bloomington, Illinois	>34,000	(1), (3)
W-186	Idem	>31,000	(2), (3)
W-96	Germantown, Ohio	>34,000	(1), (3), Leverett, 1902, p. 363.
W-152	North Hampton, Ohio	>39,000	(2), (3)
W-263	Gahanna, Ohio	>37,000	Rubin and Suess, unpublished.
W-87	Otto, New York	>35,000	(1), (3), MacClintock and Apfel, 1944.
W-100	Port Talbot, Ontario	>32,000	(1), (3)
W-194	Amber, Ontario	>34,000	(2), (3)
W-121	Toronto, Ontario	>30,000	(2), (3), Watt, 1953.
Y-269	Missinaibi River, Ontario	>29,630	McLearn, 1927.
Y-270	Idem	>30,840	Idem
Y-254	Les Vieilles Forges, Three Rivers, Quebec	>29,630	
Y-255	Idem	>30,840	
Y-256	Pierreville, Quebec	>29,630	
W-189	St.Pierre-les-Becquets, Quebec	>40,000	(2), (3)
Y-242	Idem	>30,840	
		>29,630	
W-157	Hillsborough, Nova Scotia	>40,000	(2), (3)
Y-283	Salmon Point, Prince Patrick Island, Northwest Territories	>25,300	
Y-284	Landing Lake, Prince Patrick Island, Northwest Territories	>31,840	

* W = Washington; Y = Yale.
** Numbers in parentheses refer to these titles:
(1): Suess, 1954.
(2): Rubin and Suess, 1955.
(3): Flint and Rubin, 1955.

The 25 "old" samples are listed in table 2. It is very unlikely that all are referable to a single stratigraphic horizon. All except W-99 (wood imbedded in Mankato(?) till and probably redeposited secondarily) came from

horizons overlain by glacial drift. In many instances the age of the overlying drift was not known; in a few it was believed to be Cary. However, W-87 was overlain by drift classified as lower Wisconsin and W-115 and W-139 by Iowan till.

Five of the samples (W-87, W-121, W-157, Y-250, Y-269) were underlain by drift; three more (Y-256, Y-283, Y-284) are believed, on indirect evidence, to postdate at least one glaciation. In only one instance (W-87) had an opinion as to the correlation of the underlying drift been stated; this was considered to be Illinoian(?). At no locality is the drift underlying a sample weathered so thoroughly as to compel its assignment to the Illinoian stage or an older one.

At 10 of the sample localities taxonomic identification of the samples (wood and/or peat) was made. The results show that at 7 of the 10 (W-99, W-101, W-102, W-121, W-157, Y-250, Y269-270) the ecology of the sampled material implies a climate not unlike that existing at the same locality. At 3 of the 10 (W-87, W-100, W-194) the ecology implies a climate cooler than that of today. One of the group of 7 (W-121), peat uncovered in subway excavations at Toronto, Ontario, implies a climate somewhat cooler than that indicated by the well-known Toronto interglacial beds exposed a few miles away (Flint and Rubin, 1955, p. 656).

These rather meager data compel no single stratigraphic correlation. With one exception[6] the dates of the group are older than those of samples collected from drift known to be of Wisconsin (including lowest Wisconsin) age (Flint and Rubin, 1955). Two possibilities, then, present themselves. One is that the "old" samples represent an interglacial age, most likely the Sangamon, and that the underlying drift is Illinoian or older. Against this hypothesis are the rather cool climate implied by some of the samples, the lack of evidence, in any of them, of climates warmer than the present ones, and the lack of evidence of a long lapse of time between the episodes of deposition of the underlying and overlying drift bodies. This argument is not strong, because it rests partly on negative evidence and because it can be contended that the samples collected thus far happen to represent very early or very late, and hence cool, parts of an interglacial age.

The second possibility is that the samples represent an interval (or intervals) that was ice-free at the localities of their occurrence, that antedates all of the Wisconsin drift of the literature, but that is not as old, and perhaps was not as long, as the Sangamon interglacial age. This less conventional hypothesis apparently is not excluded by the facts known at present. It seems to be more nearly compatible with the data outlined here than is the alternative, and it receives at least some support from the stratigraphic relations exposed at Sidney, Ohio, where Wisconsin till, dated at slightly less than 23,000 yr B.P., overlies a till in which a soil is developed (Flint and Rubin, 1955, p. 651).

As yet we can not choose between these possibilities. It is tempting, however, to speculate on the consequences of accepting the second one, if new

[6] W-67 and W-186, from the same locality, were imbedded in till considered, before dating, to be Tazewell.

evidence should increase its probability. The concept of a post-Sangamon climatic sequence that consisted of one or more glaciations, separated from a succeeding Wisconsin glaciation (beginning prior to 25,000 years ago) by a cool and perhaps long interval of glacier recession, is not less compatible with the drift sequence in the Cordilleran region than the one currently followed. In many parts of that region a deeply weathered drift is succeeded by two less-weathered drifts, one of which is distinctly fresher than the other. Conceivably only the younger of the two is correlative with the Wisconsin drift of the Central Lowland. Similarly, in Europe, the much-debated break between Würm I and Würm II, between Younger loess I and Younger loess II,[7] and between Warthe drift and Weichsel drift could conceivably be the interval of glacial recession represented by "old" samples discussed in this paper.

POSTSCRIPT

Made more accurate by improved techniques, radiocarbon dates are leading us gradually back into Pleistocene history and are helping surmount the great difficulties of correlation on ordinary stratigraphic evidence, difficulties that are far greater in terrestrial glacial deposits than in marine deposits of pre-Pleistocene age. The present paper aims at demonstrating further the usefulness of radiocarbon. The general consistency of the dates discussed, both internally and with dates published earlier, is good. Probably it is significant that most of the inconsistent dates in the present lot are on samples of shell (Y-215, Y-216, Y-233, Y-271).

This is a report of progress. The sooner it is superseded by improved data, the farther we shall be along the road of stratigraphic understanding.

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[7] See the brief discussion of W-173 in Flint and Rubin, 1955, p. 657.

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THE SEDIMENTARY ORIGIN AND STRATIGRAPHIC EQUIVALENCE OF THE SO-CALLED CRANBERRY AND HENDERSON GRANITES IN WESTERN NORTH CAROLINA

F. DONALD ECKELMANN* and J. LAURENCE KULP

ABSTRACT. The rocks along the northwest limb of the Spruce Pine synclinorium were mapped by A. Keith as Cranberry Granite, and those along the southeast limb were mapped as Henderson granite. These rocks are considered to be sedimentary in origin, because of (1) the complete lack of intrusive relations between the Cranberry or Henderson and the overlying metasediments, (2) interbedding of these rocks with metamorphics of known sedimentary origin, (3) the presence of primary composition banding in at least 80 percent of the observed outcrops, and (4) the high "rounding index" of zircon concentrates obtained from even the more homogeneous phases of granitic gneiss. The identity of the Cranberry and the Henderson is established by (1) the similar stratigraphic position of both formations at the base of the Roan-Carolina complex, (2) detailed mapping in the northwest corner of the unpublished Morganton quadrangle where Cranberry and Henderson meet and where they are found to be indistinguishable, and (3) the similarity in range of mineralogical variation in each formation. Thus the rocks are older than the Roan-Carolina complex rather than younger, as previously reported. Within the main sedimentary sequence, small crosscutting granite masses of limited extent have been found which themselves have been metamorphosed. They have been emplaced prior to the second metamorphic cycle of the region and are distinct from the Cranberry-Henderson metasediments. Zircons from ten representative samples of the various phases of the Cranberry and Henderson rocks were examined as to size, shape and evidence of rounding. These characteristics can be used to distinguish between orthogneisses and paragneisses of granitic composition. The results of the zircon study are in agreement with the field and petrographic evidence concerning the origin of these rocks.

INTRODUCTION

Over the past ten years field parties of the U. S. Geological Survey have mapped about 300 square miles of the metamorphic complex in the vicinity of Spruce Pine, North Carolina at a scale of $1000' = 1''$. This large effort was related to the variety of non-metallic mineral deposits in the area. These economic deposits occur in the Roan-Carolina metasedimentary complex, peripheral to which is the contact of the Roan-Carolina with the Cranberry and Henderson gneisses. It was felt that such detailed work in the heart of the Blue Ridge Province should be tied to the regional picture. Therefore, a reconnaissance study of the surrounding Cranberry and Henderson rocks was initiated.

Location and Regional Distribution of Rocks

Figure 1 shows the broad regional geologic picture in North Carolina and surrounding states (King, 1951). The area studied is included in the Roan Mountain, Cranberry, Mount Mitchell and Morganton thirty-minute quadrangles shown in the index map (fig. 2). A generalized geologic map of the area under consideration is shown in figure 3 (Keith, 1903, 1905, 1907). The granitic gneisses on the northwest side of the Roan-Carolina complex as well as in the Cranberry quadrangle were originally mapped by Keith (1903, 1905, 1907) as "Cranberry granite." The gneisses southeast of the Roan-Carolina complex in the Mount Mitchell quadrangle were called "Hen-

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derson granite." The relation between these two rock units in the northwest corner of the Morganton quadrangle was supplied by the present study.

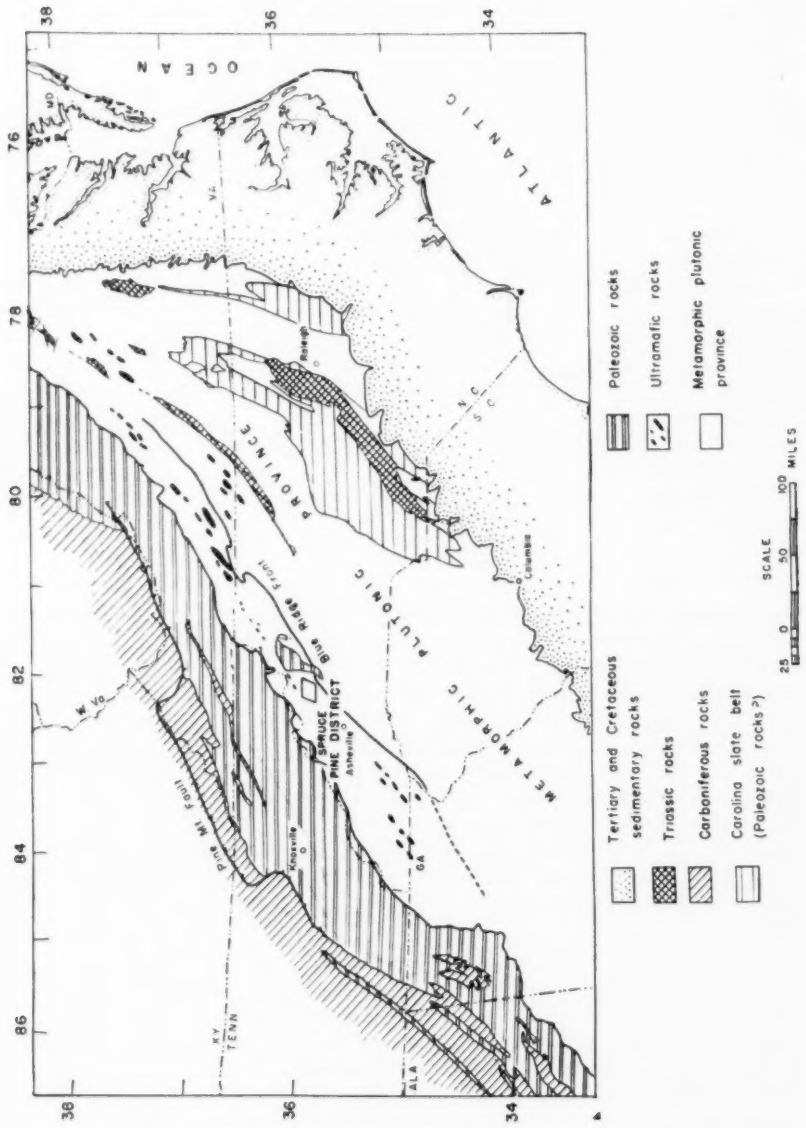

Fig. 1. Regional geology of North Carolina and adjacent states.

Background Work

The only published work on the origin and relation of the Cranberry and Henderson formations is that of Keith (1903, 1905, 1907). Keith described the Cranberry as a granite of varying texture and color in which there are schists and granitoid gneisses derived from the granite. He considered the Henderson a granite which had some schist and gneiss areas. With regard to crosscutting relationships he claimed the granite intruded the Roan and Carolina gneisses but failed to cite specific locations, saying that the contact is usually obscured by heavy forest growth and metamorphism. Both rock units were considered to be of Archean age, intrusive into the Roan and Carolina gneisses, also of Archean age.

Reconnaissance work by the U. S. Geological Survey parties from 1948 to 1952 (Brobst and Kulp, in press) produced observations that appeared to contradict Keith's conclusion. This was due in part to the superior exposures now available along numerous roads and railroads and in quarries, and to more detailed field mapping. The observations indicated that neither the Cranberry nor the Henderson are intrusive granites, but rather that they represent metasedimentary rocks which underlie the Roan-Carolina sequence, and are probably of late Precambrian age. Further, there was a strong suggestion that the Henderson and Cranberry units were stratigraphically equivalent and continuous into each other.

With this information a more detailed study of critical areas was conducted.

General Statement of Problem

This study has three main aspects: (1) The structural and stratigraphic relations between the Cranberry and Henderson rocks and the Roan-Carolina complex, (2) the relation of these two units to each other, and (3) a determination of the origin of the Cranberry and Henderson rocks immediately underlying the Spruce Pine synclinorium.

The results of such a study have considerable importance for the regional geology of this part of the Blue Ridge Province and the economic geology of the Spruce Pine District. Both the Roan-Carolina and the Cranberry-type rocks are widespread in the region. Clearly a late Precambrian metasedimentary sequence provides an entirely different basis for geologic interpretation than a very ancient metasedimentary sequence intruded by granite batholiths. These alternate premises greatly affect one's conclusions as to the origin of the pegmatite, alaskite, and ultramafic bodies in the Spruce Pine District.

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ROCK DESCRIPTION

Carolina Gneiss

The rocks of the Carolina gneiss include all the various mica gneisses and mica schists in the Spruce Pine District. These rocks are interlayered with those described below as Roan gneiss. Moreover, the two types of rocks grade into each other along and across strike. A detailed description of these two rock types and a discussion of their origin and relationship to each other will be found elsewhere (Brobst and Kulp, in press). The conclusion may be summarized by saying that the Roan-Carolina sequence appears to be a single sedimentary unit with varying lithology, complicated by intense isoclinal folding and plastic deformation.

The Carolina formation consists of mica gneiss and mica schist which grade into each other. However, there are definite criteria for distinguishing the rocks most characteristic of each unit. Grain size generally ranges between 0.2 and 0.5 mm in the gneiss and between 0.5 and 5 mm in the schist. Foliation is more marked in the schist, chiefly because it contains more mica. Muscovite generally dominates over biotite in the schist, but there are exceptions, especially in the transition zones between the gneiss and the schist. The gneiss is more homogeneous than the schist, in which thin layers of mica separate streaks and pods of feldspar and quartz; snowball garnets are also more common and more abundant in the schist. Texturally both rock types may be granoblastic; but augen, flaser, and cataclastic effects, such as crenulated and broken mica and distorted metacrysts with the foliation wrapped around them, are more common in the schist.

Roan Gneiss

The Roan gneiss consisting of undifferentiated hornblende rock contains many kinds of hornblende gneiss and schist, together with layers characterized by actinolite, diopside, epidote, zoisite, and garnet. These rocks are interlayered with the mica rocks and are transitional into them. Nowhere in the Bakersville-Plumtree area were hornblende rocks found to crosscut other rocks. The individual layers range in thickness from less than an inch to several tens of feet. The grains vary in size from 0.1 mm to 3 mm.

The rocks of the Roan gneiss range in color from gray-green to dark greenish black, depending on the relative abundance of light and dark minerals. Foliation is generally well marked, because of the alignment of the hornblende or other needle-shaped or platy minerals. The structures are schistose to gneissic, and the textures are granoblastic; some layers show evidence of deformation after recrystallization.

The major mineral constituents of the Roan are hornblende, actinolite, oligoclase, andesine, quartz, epidote, zoisite, garnet, diopside, chlorite, and biotite. The accessory minerals include sphene, rutile, apatite, zircon, allanite, sericite, calcite, orthoclase, magnetite-ilmenite, pyrite, and pyrrhotite.

In weathering, the hornblende rocks have a tendency to break off in slabs. The soil developed from them is generally dark red to reddish brown and is the most plastic of the soils in the area. It is unlike that derived from the mica rocks in that it contains less quartz and little or no bleached mica.

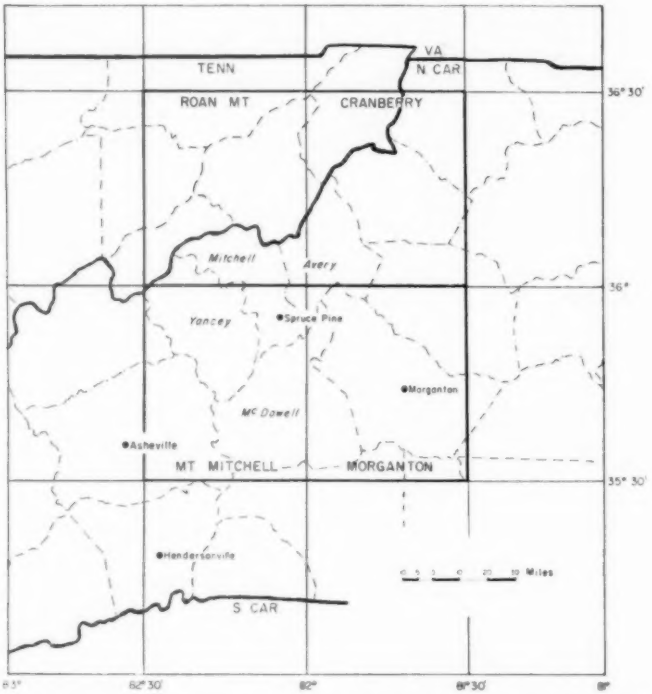


Fig. 2. Index map of western North Carolina and eastern Tennessee. The four 30″ quadrangles in which reconnaissance work was done are bounded by a heavy black line.

Cranberry and Henderson Rocks

Field classification.—The Cranberry and Henderson rocks, which show a similarity in range of mineralogical variation, may be megascopically separated into three distinct groups. The first group is characterized by primary compositional bands bounded by sharp planes. The bands range from a few inches to several feet in thickness. The more feldspathic bands generally lack foliation, whereas the more micaceous bands always possess foliation. This group comprises the bulk of both formations. The second group lacks banding but has well developed foliation. The third group includes only massive rocks, void of both banding and foliation.

The general appearance of both the coarse and the more finely banded rocks differs primarily in thickness of the individual bands. In outcrops measuring in tens of feet, the individual bands are found to be continuous throughout the exposure and of uniform thickness even where the bands are less than an inch thick. Each band exhibits sharp contacts with the rock above and below. Both characteristics are seen in a road cut at the junction of Millers Branch and Brummett Creek 0.5 of a mile north of Webb. The larger percent of this outcrop consists of coarsely banded, light-colored quartzo-

feldspathic rock with thin interbedded bands of mica schist. Small individual micaceous bands, varying in thickness from 2 cm to .2 cm, also exhibit sharp contacts, uniform thickness, and continuity throughout the outcrop. Foliation planes developed in the schist bands parallel the bedding while feldspar and quartz porphyroblasts have developed in the coarser light-colored bands.

Exposures illustrating the finer-banded rocks are seen in a new road cut 3.2 miles east of Minneapolis (on the road to Newland) and 1 mile southwest of Newland on the Cow Camp road (to Mount Pleasant). The second of these localities is a continuous half-mile outcrop of finely banded Cranberry gneiss interbedded with hornblende gneiss (pl. 3). This same fine banding can be duplicated in the area mapped as Henderson granite by Keith. Two such examples are found in road cuts 7 miles south of Old Fort on state highway 64 and 0.3 mile north of the junction of routes 26 and 26A on 26.

Illustrations of the more coarsely banded rocks are found in a railroad cut 0.5 mile northwest of Relief along the North Toe River (table 1), 0.2 of a mile west of Newland (on the road to Minneapolis), and 0.6 of a mile southwest of Old Fort on route 70. The first two examples are in areas mapped by Keith as Cranberry granite, and the third is in an area mapped as Henderson granite.

The second rock type, possessing only foliation, is best seen at three localities: (1) a quarry northwest of Crossnore, (2) the east end of Doe Hill, and (3) a new road cut immediately north of Relief on the road to Pigeon Roost.

The Crossnore quarry, located 0.7 of a mile northwest of Crossnore, on both sides of the road between Crossnore and Mount Pleasant, contains a massive foliated granitic rock of medium to coarse texture. The main exposure is 500 feet long. The south end of the outcrop is interrupted by a small stream valley. Beyond the alluvial valley floor is normal finely banded quartzo-feldspathic Cranberry gneiss. It is therefore not possible to observe the field relations between the massive granitic rock of the quarry proper and the finely banded rocks farther south. North of the massive rocks found in the quarry lie well banded mica gneiss and schist and light-colored quartzo-feldspathic rock identical with the Cranberry south of the stream valley. The intermediate contact area unfortunately is covered. A secondary quarry located on the east side of the road contains rocks mineralogically identical to the main quarry except for a more pronounced foliation parallel to prominent near-horizontal joint planes, not developed in the first quarry, and microscopically observable cataclastic texture. The attitude of the foliation at the north end of the second quarry is variable. Again, the contact area separating the banded Cranberry gneiss to the north from the massive foliated granitic rocks of the quarry is covered.

The abrupt change in rock types to either side of the covered contact zone favors an intrusive origin for the foliated granitic gneiss of the quarry. Zircon work, reported in detail below, strongly supports this mode of origin.

The east end of Doe Hill and exposures north of Relief are examples of massive foliated granitic gneisses which in hand specimen and thin section are identical to the Crossnore Quarry rock.

TABLE 1*

Cranberry Gneiss Section at Relief

Location: Base of section begins .25 mile north of Toe River bridge (on the east side of the river) at the town of Relief.

Geology: The Cranberry gneiss here is very well banded, striking approximately N 60 E, dipping 40-70 SE.

Feet	Inches	Top of the section
3		Numerous small lenses of plastically deformed quartzofeldspathic gneiss separated by thin layers of biotite schist.
15		Fine-grained white and gray layers of feldspathic and quartzose gneiss.
1		Porphyroblastic gray gneiss (porphyroblasts are ¼″ across) with fine biotite foliation wrapping around the porphyroblasts.
5		Thinly banded white to light gray feldspathic and quartzose rocks with some fine-grained biotite present.
9		Fine-grained, blue-gray to black biotite and hornblende schist.
6		Dark biotite-rich layers (with some hornblende) grade back and forth across strike to lighter porphyroblastic layers containing thin streaks of added quartz and feldspar.
	4	Foliated, white quartzofeldspathic rock with porphyroblasts of quartz and feldspar up to ⅜″ across with foliation wrapping around them.
1	5	White to gray gneiss, composed mainly of quartz and feldspar, with thin irregular streaks of fine biotite. Grades into a fine-grained biotite schist in the upper 6″.
	4	White quartzofeldspathic gneiss with up to 2% fine biotite.
7	6	White to gray feldspar and quartz gneiss with porphyroblasts ¼″ in diameter of the same composition. Thin irregular layers of fine biotite are present. The top six inches is biotite schist. (Note the same sequence in the 1'5″ unit above).
	10	Foliated, highly feldspathic layer with 2-3% biotite.
	6	Fine-grained, dark gray phyllitic rock.
	5	White quartzite.
5		Thinly banded feldspathic and quartzose rocks.
	4	Brown biotite schist.
	4	Brownish-white feldspar gneiss.
	3	Brown biotite schist.
3		Nine white to gray bands of feldspathic rock. The percent of biotite present varies, being higher in the darker layers.
10		Porphyroblastic, dark gray, feldspathic quartz gneiss with fine-grained biotite. The porphyroblasts are as much as ¼″ in diameter. Some thin irregular streaks composed wholly of quartz and feldspar are present.
2		Foliated, white, feldspathic quartz gneiss.
7		Porphyroblastic (¼″ diameter), white, quartzofeldspathic gneiss with irregularly distributed streaks of fine biotite which becomes thin continuous layers up to 3″ thick near the top of the unit. The dark material forms 25% of the unit.
1		Brown biotite schist.

* Compiled from data in a written communication received from Donald A. Brobst.

TABLE 1 (Continued)

Feet	Inches	Top of the section
3		Foliated, dark-colored, biotite amphibolite layers containing light-colored pods of quartz and feldspar that make up 30% of the total rock.
6	6	Foliated, porphyroblastic, dark-colored amphibolite biotite gneiss. Numerous thin, discontinuous feldspar quartz layers. The biotite has probably formed at the expense of the amphibole.
3	6	Massive to porphyroblastic, foliated, white feldspathic quartz rock with some amphibole and biotite, both minerals being accessories. The porphyroblasts are ⅜-½" in diameter.
5		Fine-grained, variable light gray to blue-gray friable phyllite.
18		Foliated, porphyroblastic (¼" diameter), light-colored gneiss containing thin dark amphibolite layers (½-2" thick) which form 20% of the unit.
2		Foliated dark brown schist which passes laterally into less schistose material in the upper parts of the outcrop.
2		Porphyroblastic (⅛" diameter), white, feldspathic quartz gneiss.
	7	Dark brown biotite schist.
2	6	Porphyroblastic, brownish-gray feldspathic gneiss with thin streaks of quartz and feldspar.
4		Porphyroblastic (¼" diameter), brownish-gray feldspathic gneiss. Porphyroblasts are of feldspar.
		Bottom of section. The section could be continued above and below what is given here, but is monotonously similar to that already quoted. This outcrop of considerable length is one of the best exposures of well banded, foliated, layered Cranberry gneiss.

The third rock group includes medium- to fine-grained, equigranular granitoid rocks with no evidence of foliation. The extent of individual outcrops in some cases are known to be small, usually being measured in tens of feet. No outcrops were found where crosscutting relations between this rock type and the banded Cranberry could be seen, though the character of the rock suggests that such relations may exist and could be found by more detailed field work in areas such as the south slopes of Fork Mountain. An outcrop of this rock type is seen 3.1 miles south of Iron Mountain Gap on route 26. Here one finds a 50-foot exposure consisting of a medium-textured, equigranular, feldspar-quartz rock which shows no banding or foliation in the central portion of the outcrop. Toward one end of the outcrop faint foliation is developed while at the very end of the outcrop distinct banding can be seen. The attitudes of the foliation and the well developed banding are essentially the same. A second example of this rock type is found 0.1 mile northwest of Elk Park on route 194. Contact relations between the banded Cranberry and the massive rock are obscured because of intense shearing. More extensive outcrops of this rock type are found only on the southwest slopes of Fork Mountain. Contact relations here are obscured because of talus material and heavy forest growth. In the few areas where this rock type is found the occurrence might be explained by intense local migmatization and feldspathization resulting in a very light-colored porphyroblastic rock possessing little or no foliation.

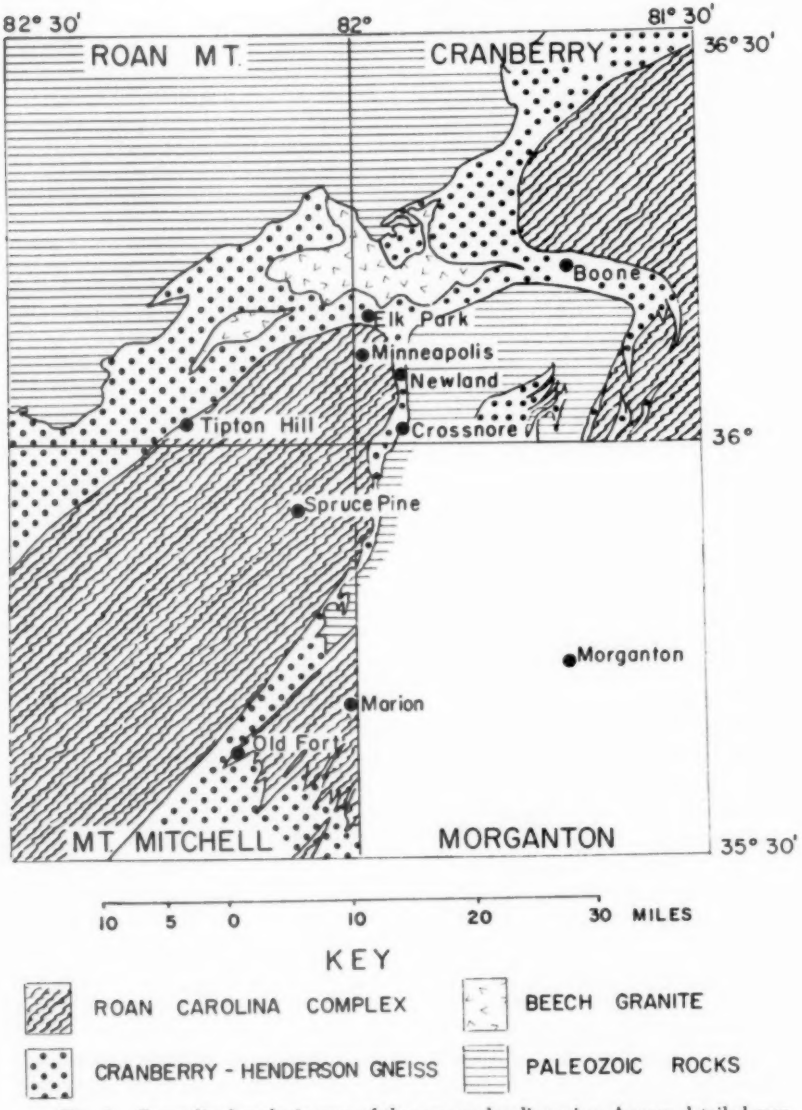


Fig. 3. Generalized geologic map of the area under discussion. A more detailed map of the northwest corner of the Morganton quadrangle is given in figure 9.

Frequency of occurrence of various phases.—An attempt was made to evaluate the relative quantities of metasedimentary rocks as against the more massive foliated and non-foliated gneisses. This was derived from hundreds of

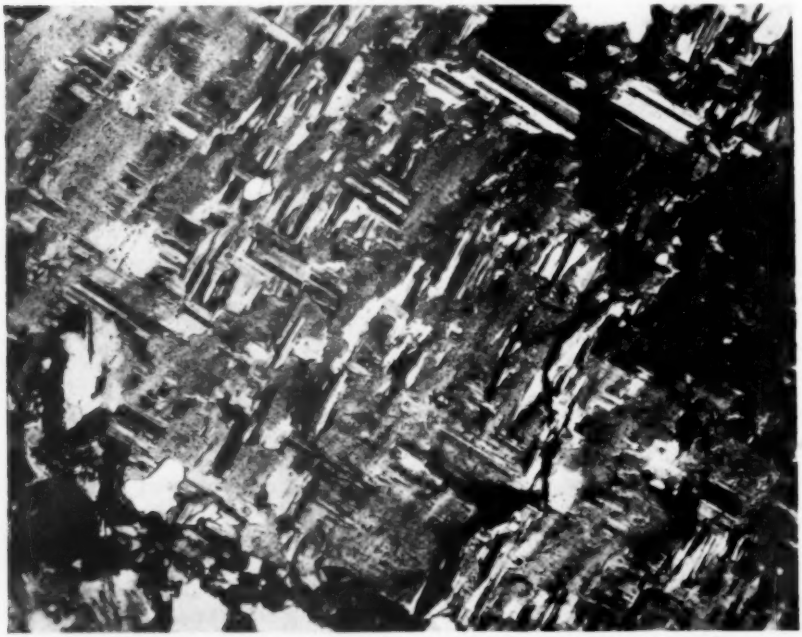
outcrops observed in the areas exposed along the limbs and at the nose of the Spruce Pine synclinorium. A crude estimate puts the amount of metasedimentary rocks (clearly banded) exposed in the area mapped as Cranberry or Henderson at over 80 percent, the coarse foliated rocks less than 15 percent, and the massive medium-fine-grained rocks at less than 5 percent. Since much of each of the two latter groups may be metasedimentary, it is clear that the Cranberry and Henderson should no longer be considered intrusive granites. The names Cranberry gneiss and Henderson gneiss are more properly used for the quartzofeldspathic rocks immediately underlying the Spruce Pine synclinorium.

Petrography.—Microscopic studies indicate minor compositional differences and no distinct textural differences between the main phases of the Cranberry and Henderson gneisses. For this reason the following petrographic description will apply to both units with note being made of minor differences observed. Plates 1 and 2 show four photomicrographs of different varieties of the gneisses.

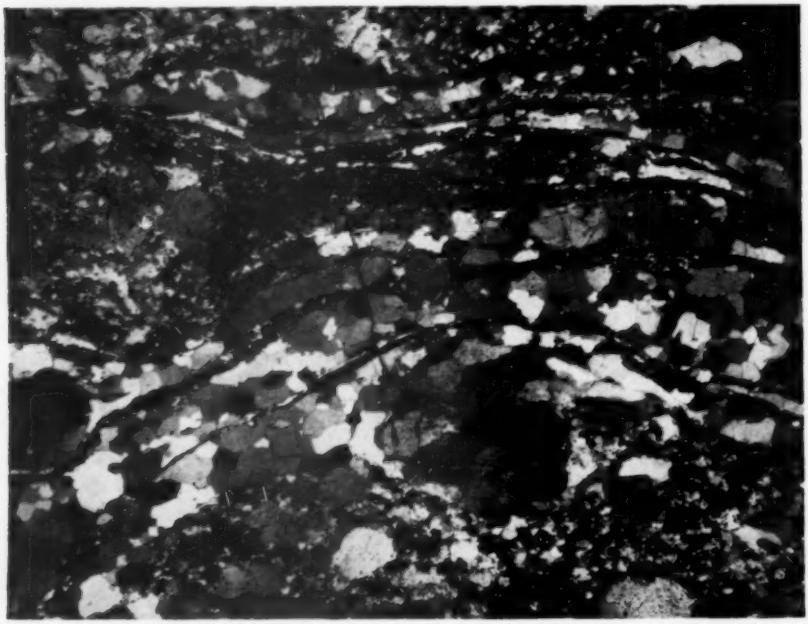
The main phase of the Cranberry and Henderson gneisses is a coarsely banded porphyroblastic granitic gneiss with a fine-grained groundmass. Local lenses and bands of mica gneiss and schist are found throughout the entire section and are more common as the contact with the overlying Roan-Carolina complex is approached. Microcline, plagioclase (albite or oligoclase), quartz, and biotite are the major constituents in all rocks examined. Minor constituents include augite, chlorite, epidote, clinozoisite, hornblende, tourmaline, sericite, and muscovite. Larger percentages of sericite and hornblende are found in the Cranberry and of epidote, clinozoisite and muscovite in the Henderson. Hornblende is encountered more often (though not in higher concentrations) in the Cranberry, muscovite more often in the Henderson, and augite and tourmaline exclusively in the orthogneiss phases of the Cranberry. Accessory minerals include allanite, apatite, calcite, fluorite, garnet, leucoxene, opaque iron ores, rutile, sphene, and zircon. Calcite, fluorite, and garnet are found only in the Cranberry while all other accessory minerals are common to both units.

Porphyroblasts of microcline, perthite, and plagioclase embedded in a fine-grained, cataclastic groundmass characterize the great majority of rocks examined. The porphyroblasts are 1-4 mm in diameter while individual crystals in the groundmass are usually less than 0.5 mm. Occasionally a complete transition in grain size exists between porphyroblasts (1-2 mm) and the fine-grained groundmass (1 mm or less in diameter). Foliation is present in all rocks while banding is observed in only a few thin sections of the Cranberry. Granulation, shearing and rotation of porphyroblasts is common to all rocks, though varying considerably in degree from one area to another. Where this action was most pronounced only microcline and perthite remain as large grains and are themselves often fractured. Polysynthetic twins in albite and oligoclase often show considerable bending without fracturing. Foliation is due to streaks of chlorite, biotite, muscovite and sericite, with epidote, clinozoisite, and apatite sometimes present. Occasionally finely granulated streaks of quartz and feldspar, or hornblende and biotite, are responsible for

PLATE 1

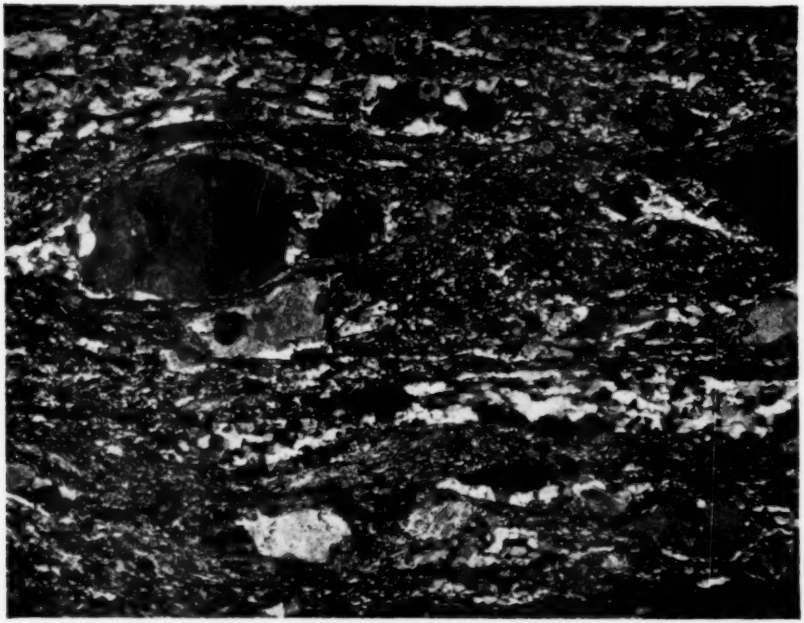


A. Patch perthite in Cranberry orthogneiss from a quarry 0.7 mile northwest of Crossnore. 11-52. Crossed nicols. ×50.



B. Quartzofeldspathic Cranberry paragneiss with granulitic texture, from 1.7 miles southwest of Newland along the Cow Camp road. 64-52. Crossed nicols. ×15.

PLATE 2



A. Granulated quartzofeldspathic Cranberry paragneiss, 0.3 mile west of Newland. 19-52. Crossed nicols. ×20.

B. Porphyroblastic, granulated Henderson paragneiss, 6.4 miles south of Old Fort on the road to Wild Cat Knob. 55-52. Crossed nicols. ×15.

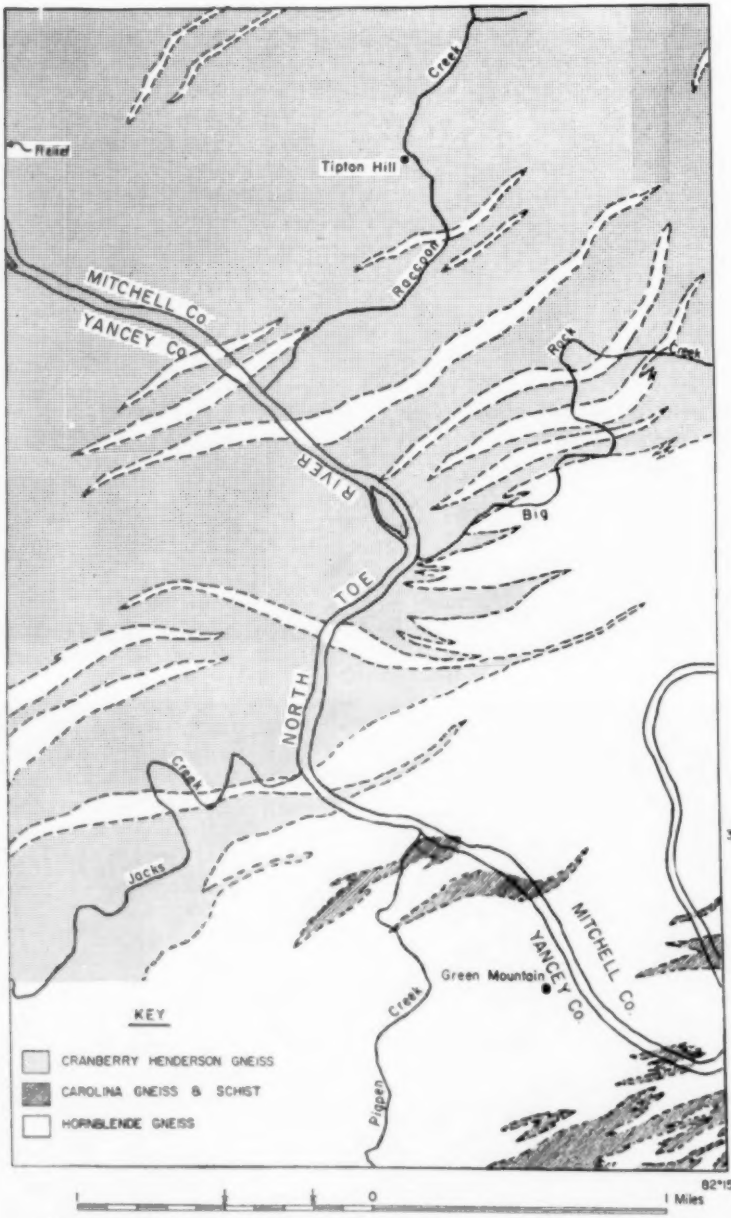


Fig. 4. Geologic map of the area between Green Mountain and Pigeon Roost. The gradational contact between the Cranberry and Henderson gneisses and the Roan-Carolina complex of the Spruce Pine District is demonstrated.

the foliated appearance of a rock. A few specimens show no development of porphyroblasts, being even grained throughout. Much of the microcline present exhibits no grid structure and can be easily mistaken for orthoclase. Inclusions of sericite, epidote and apatite are common in the porphyroblasts. When fractured and broken the microcline will sometimes show healing by quartz and feldspar or develop interstitial quartz, mica and feldspar. Perthite is ubiquitous and varies in degree of development from string through film, vein, and patch varieties. Patch perthite, though well developed, is found only in two specimens, one of which is unquestionably an orthogneiss (Crossnore quarry). Twinned microcline is also common in the groundmass. Occasionally microcline and plagioclase contain inclusions of epidote, quartz and zircon. Quartz is present mainly in the groundmass. Biotite is normally dark brown and pleochroic but in several rocks is a very dark green variety. Epidote is fine grained, prismatic or rounded, and aggregates in clusters and streaks or is scattered through a thin section. Chlorite in a few grains has an anomalous Berlin blue color under crossed nicols. Green aluminous augite was found only in the Crossnore quarry sections. Sphene occurs as wedge-shaped crystals, drawn-out fine-grained clusters, or as a rim about ilmenite.

STRATIGRAPHIC RELATIONS OF THE HENDERSON AND CRANBERRY

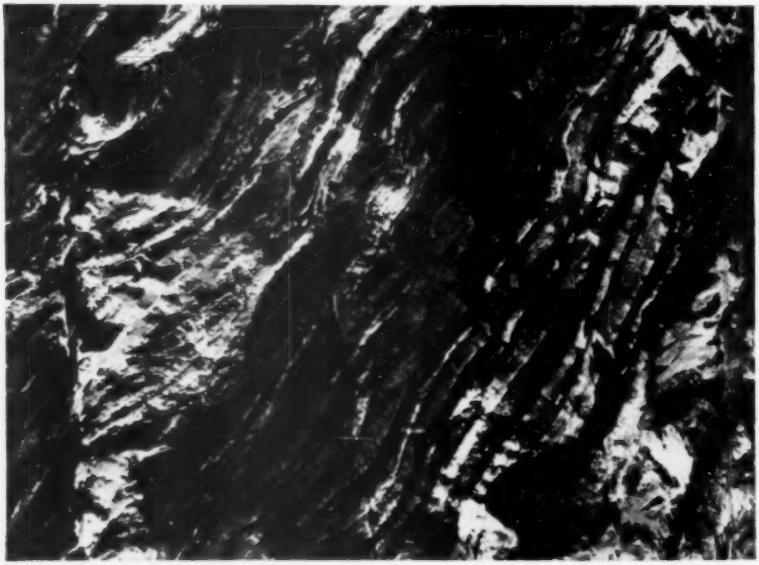
Nature of Contacts with Overlying Metasedimentary Rocks

The interbedding of the Cranberry and Henderson formations with the overlying metasedimentary rocks is the most striking single feature observed at each contact zone studied. In the case of the Cranberry formation, this is best illustrated at three localities: (1) a road cut 3.2 miles east of Minneapolis (pl. 3), on the road to Newland, (2) a continuous half-mile outcrop beginning 1.7 miles southwest of Newland along the Cow Camp road, and (3) along N. C. 197 between Green Mountain and Relief (fig. 4). Similar examples in areas mapped as Henderson granite by Keith were found. The best of these is seen along a 2-mile railroad cut between Jackson Knob Loop (at the head of Honeycut Creek) and the town of Ashford.

In each case the alternating bands of granitic gneiss and mica and hornblende schists are continuous throughout the outcrop and of nearly uniform thickness. No evidence of chilled margins, baked contacts, apophyses, or inclusions of foreign material has been found in the granitic gneiss bands as would be expected if the banding were a lit-par-lit phenomena.

Traverses were made across the contact zone between the Cranberry and overlying rocks in several areas for the purpose of determining the character of the transition zone between the two rock types. One area considered was a nearly continuous road cut 1.0 mile long on the northwest slopes of Spanish Oak Mountain southwest of Newland. A second transition zone studied is exposed along the Blue Ridge escarpment between Humpback Mountain and Ashford. In both cases the interbedded character of the adjacent rock types was as described above. The transition zone along Spanish Oak Mountain is between the Cranberry formation to the north and a thick amphibolite unit in the Carolina schist to the south. Going southward along the Spanish Oak Mountain section, the percent of Cranberry type bands decreases along with

PLATE 3



Photograph of the transitional contact between Cranberry gneiss and the Roan-Carolina complex. The roadcut is 3.2 miles east of Minneapolis on route to Newland. The interbedding of the Cranberry gneiss with the overlying Roan-Carolina complex is typical of the transitional zone between the two rock units.

a corresponding increase in the percent of amphibolite bands. Thin bands, identical with those in the main Cranberry mass to the north, persist into the amphibolite as much as 0.5 mile beyond the mapped contact in this area of Carolina schist and Cranberry gneiss. A detailed measured section along a portion of this transition zone is graphically represented in figure 5, and illustrates the gradual transitional character of the contact zone.

In these two areas and others studied less intensively, a consistency of strike exists throughout the transition zone.

Stratigraphic Equivalence of the Cranberry and Henderson Gneisses

As mapped by Keith, a belt of outcrop 2 to 3 miles wide of Henderson extends along the base of the Blue Ridge Escarpment between Old Fort and Marion. At a point on highway 26, .2 mile north of the junction between 26 and 26A, this same band of Henderson narrows to less than a half mile in width and continues to the northeast corner of the Mount Mitchell quadrangle. A similar belt of outcrop 1 to 2 miles wide of Cranberry extends to the southwest corner of the Cranberry quadrangle. Therefore any surface connections between the Cranberry and Henderson formations will be found in the northwest corner of the Morganton quadrangle (fig. 3), an area which has not been previously reported on in the literature.

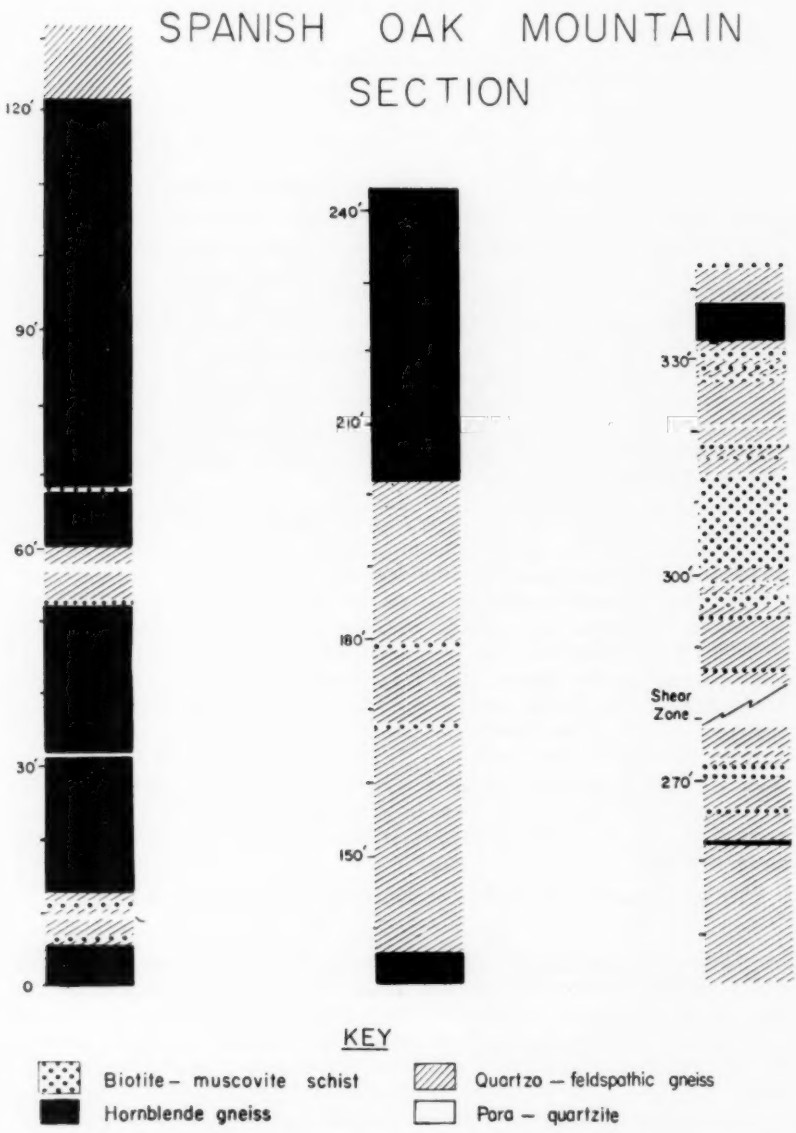


Fig. 5. Stratigraphic section at Spanish Oak Mountain. The measured section demonstrates the transitional contact between the Cranberry gneiss and the Roan-Carolina complex of the Spruce Pine District.

Geologic mapping of this critical area (fig. 9) clearly demonstrates:

(1) the Cranberry and Henderson formations are continuous into each other,

(2) the attitude of these units is consistent with the structure of the adjacent overlying Carolina gneiss and schist to the west, and north,

(3) interbedding between the Carolina rocks and underlying granitic gneisses exists in a wide transition zone between the two rock groups, exposed in the Blue Ridge Escarpment.

In each traverse across the critical area, bounded essentially by the Blue Ridge Parkway and route 221, a gradual transition exists from Carolina rocks with no included bands of granitic gneiss, through a zone where bands of both types were interbedded, to a section consisting completely of Cranberry-Henderson rock types (lowest stratigraphically). The width of outcrop of this final zone is approximately 1.0 mile at North Cove, narrowing to 0.5 mile, 2 miles south of Chestoa. This same width continues to Chestoa where the section gradually becomes 1.5 miles wide to the north on passing into the Cranberry formation mapped by Keith in the southwest corner of the Cranberry quadrangle. All variations of rock types found here can be duplicated by specimens collected in the main areas mapped by Keith as Cranberry and Henderson granite.

Banding in Main Masses of Henderson and Cranberry

The characteristic banding of the Henderson and Cranberry described above is not restricted to the border zone between these units and the overlying rocks. The approximate abundance of banded rock as compared with abundance of foliated and massive rock has been given above. From these estimates it is evident that the great majority of Cranberry and Henderson outcrops exhibit this feature. It is thus possible to find extensive areas of highly banded rock in the areas mapped by Keith as intrusive granite. Two examples of this in the Henderson granite are: (1) 8.8 miles south of Old Fort on the road to Bat Cave, and (2) 0.6 miles southwest of Old Fort on route 70. Similar exposures in the Cranberry are found along the entire 4.5-mile railroad cut between the towns of Green Mountain and Relief. A second example in the Cranberry formation is found along Griffith Branch Creek north of Webb.

MICROSCOPIC STUDY OF ZIRCONS FROM CRANBERRY AND HENDERSON ROCKS

General Background

During metamorphism a rock can be so thoroughly reconstituted that all its original features may be lost. Field evidence alone may then be insufficient to establish the original character of the rock. Thus genetic problems of gneisses and schists, especially of granitic composition, commonly require special methods of investigation additional to the usual detailed field observations.

Peripheral to the Spruce Pine area banded granitic gneisses of debatable origin abound, and hence it has been necessary to study the size and shape of zircon concentrates from these rocks for additional clues to their origin. Such a study is based on the following assumptions. During transportation

zircon grains will be abraded and gradually become more and more rounded (Smithson, 1939, 1941; Tyler and others, 1940; Armstrong, 1922; Poldervaart, 1955). In igneous rocks the majority of zircons will be euhedral with only a small (but variable) percent of rounded grains present (Winchell, 1914; Groves, 1930; Poldervaart, 1955). These rounded zircons may be due to irregularities in crystal growth, disordered crystal lattices, or assimilation of country rock that contained rounded zircons. Finally, zircon being one of the most refractory minerals (Shand, 1927), it will not easily recrystallize during metamorphism (Poldervaart and von Backström, 1949).

Specific Application to Determining the Genesis of Granitic Gneisses

For many years the rounding of zircons has been used as a means of determining the origin of foliated rocks (Winchell, 1914). In recent years the term abrasion index (Allen, 1944) has been used to represent the percent of zircon crystals present in a given rock that show no evidence of rounding. Thus an intrusive granite will have a high abrasion index and a pelitic sediment a low abrasion index. This choice of terminology is unfortunate as abrasion index indirectly suggests a meaning contrary to that intended for the term. For this reason the authors propose the term *rounding index* which represents the percent of crystals showing any signs of rounding. A rock with

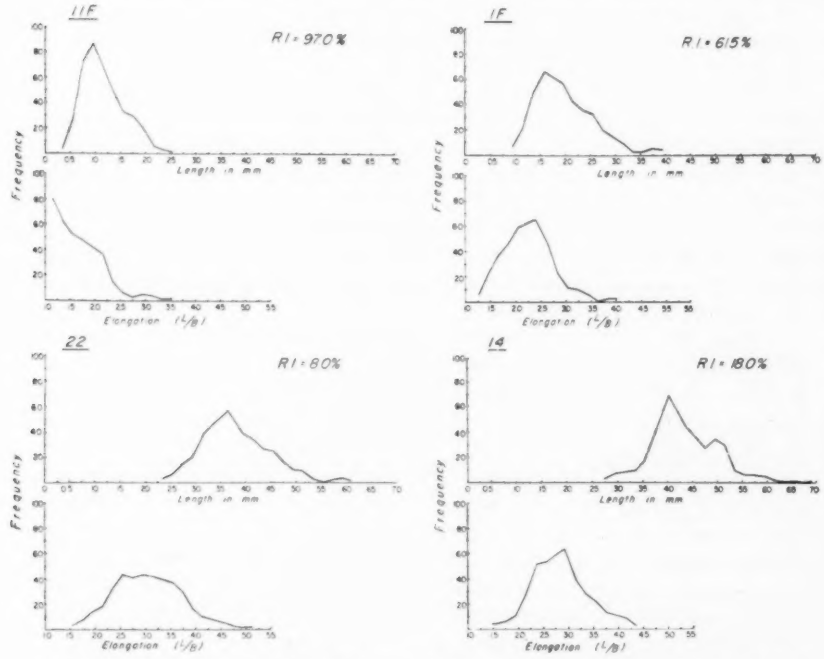


Fig. 6. Elongation- and length-frequency diagrams for zircon concentrates from rock samples 1F-52, 11F-52, 14-52, and 22-52.

a high rounding index will then contain a large percent of crystals showing some signs, however slight, of rounding due to abrasion. The term in no way indicates the degree of rounding.

Elongation-frequency and length-frequency curves (Smithson, 1939) are used in collaboration with the rounding index value to determine the origin of any given rock. Poldervaart and von Backström (1949) have pointed out that granitic gneisses of igneous origin normally have elongation-frequency maximum above 2.0, whereas paragneisses have a value below 2.0. In most cases where this maximum is below 2.0, nearly all the zircon grains show rounded terminations and usually the majority of grains are well rounded. The length-frequency maximum is also important to consider. In siltstones this value generally is 0.05-0.10 mm, whereas in sandstones and arkoses the maximum falls in the range 0.10-0.15 mm. (Poldervaart, 1955), but higher values may be found in locally derived arenites and rudites.

Interpretations based on statistical zircon studies must take into account the original habit and character of the crystals studied (Poldervaart and von Backström, 1949). Sediments derived in one cycle of erosion and deposition from a granite containing relatively short zircons will have a lower elongation-frequency maximum than sediments similarly derived from a granite containing long pencil-shaped zircons (Poldervaart, 1955). It is possible also for a sediment containing slightly rounded pencil-shaped zircons to have a higher elongation-frequency maximum than a granite containing euhedral zircons of similar length because of the difference in width of the zircons in each rock. Provided the width difference was great enough, one could have a granite pluton containing zircons twice the length of those in the adjacent metasediments, whose elongation-frequency maximum only approximately equals that of the sediment (table 2, compare samples 1F-52 and 22-52). In considering the significance of length-frequency maxima it is important to note that the length of a zircon in a sediment depends on the original length-width of the crystal, distance of transport, and number of sedimentary cycles through which it has passed.

From these considerations it is clear that the rounding index gives the most important single clue to the origin of the rock while the elongation-frequency and the length-frequency shed much light on the details of origin. A high rounding index is strong evidence of sedimentary origin for the rock in question (Poldervaart, 1955). If the length-frequency and elongation frequency maxima of a sedimentary rock are both high, the rock material is probably of local derivation, and if both are low it has probably experienced long transport. A low length-frequency maximum and a high elongation-frequency maximum will be due to the small width of the zircons in question and suggests a sorting out of small thin zircon crystals from those of larger dimensions. A low rounding index is strong evidence of an igneous origin for the rock in question (Poldervaart, 1956), and variations in length-frequency and elongation-frequency maxima represent variations in habit of the original crystals.

Method

The rock samples selected for zircon study fall broadly into two groups. First, those for which field relations suggest a sedimentary origin (1F-52, 11F-52, 19F-52, 33F-52, 55F-52, 123F-52, 134F-52). Here specimens were used from exposures of Cranberry and Henderson gneiss which exhibited well defined continuous banding, considered to represent primary bedding. Second, rocks which in hand specimen and in outcrop are of doubtful origin (14-52, 22-52, 104-52).

Sample size was approximately 250 to 300 grams. After crushing, it was passed through an 80 mesh sieve (0.22 mm), caught on a 100 mesh sieve (0.16 mm), always taking care to retain the fine material which passes through the 100 mesh sieve. This results in a coarse (80-100 mesh) and a fine (greater than 100 mesh) sample. The coarse fraction is then washed in water to remove rock flour, dried and added to a 500 cc separatory funnel approximately half full of tetrabromoethane (sp. gr. 2.96). A glass stirring rod is used to agitate thoroughly the crushed rock in the heavy liquid, thus facilitating separation of the heavy and light fractions. It is advisable to stir the light fraction repeatedly, usually at intervals of 30 minutes. The time interval allows the heavy and light fractions to separate while the repeated stirring frees any heavy minerals retained in the light fraction. A glass rod with a paddle-shaped end can be slowly put down through the light fraction to the heavy fraction, and there rotated rapidly between the palms of one's hands. This agitation of the heavy fraction allows any light minerals carried down by the heavy fraction to separate out without agitating the already separated light minerals. Such a procedure saves considerable time over a repeated separation of the heavy fraction in a second separatory funnel and also makes for less loss of rock material due to handling.

After a clear separation has been attained, the heavy fraction is filtered off, washed in alcohol, and dried. The sample is now ready to be separated into a magnetic and a non-magnetic fraction. In this work a Frantz isodynamic separator was used. The conventional procedure as outlined in the instructions issued with the instrument was followed with a few modifications. At the outset, to obtain a separation of the most magnetic material, a forward inclination of 35° and a side inclination of 15°, with the magnet set at 15 amps, is used. The magnetic material obtained from this run is set aside. The chute is then tipped forward at 25° and sidewards at 15°, the magnet set at 15 amps and the non-magnetic fraction obtained above is slowly passed through the separator. Two fractions result, one being non-magnetic and the second slightly magnetic. This step is repeated several times with the newly obtained magnetic fraction, thus obtaining a maximum yield of non-magnetic heavy minerals. The non-magnetic fraction is now passed through the separator in the same manner, giving a final clean separation. The non-magnetic fraction can then be mounted in balsam on a regular glass slide with cover glass. Experience has shown that the absence of zircons in the non-magnetic coarse fraction is no assurance that the sample does not contain zircons. All the zircons may be concentrated in the finer fraction. In this work, both coarse and fine fractions of each sample were prepared and mounted separately.

A mechanical stage was used in measuring and counting individual zircon crystals observed along regularly spaced traverses. Particular care was taken to measure only unbroken crystals. Elongation-frequency curves and length-frequency curves (Smithson, 1939), were prepared from the data, using a moving average of four for all samples except DE-55F where a moving average of eight was used. The method outlined by Smithson was used in calculating frequencies, but no corrections were made as all unbroken crystals in the field of the microscope were measured along traverses spaced at regular intervals over the entire slide. If the crystals measured are only those intersecting a fixed point in the microscope field, then a correction factor must be applied, as the larger crystals will preferentially intersect the point. Throughout, 200 crystals were measured for each sample. Poldervaart and von Backström (1949, fig. 8) showed that measurement of more than 200 crystals does not change the elongation-frequency curve significantly. The proportions of euhedral and rounded zircons were also determined for each sample. When the ratio of rounded to euhedral, or vice versa, was less than 9:1, 200 crystals were counted; otherwise only 100 grains were considered.

Results

Table 2 is a compilation of all zircon data and will be a useful point of reference for the reader in the following discussion. Six of the specimens studied (1F-52, 11F-52, 14-52, 19F-52, 22-52, 104-52) were collected from the area mapped as Cranberry granite and gneiss (Keith, 1903, 1905, 1907), and four samples (33F-52, 55F-52, 123F-52, 134F-52) came from the area mapped as Henderson granite and gneiss (Keith, 1905). One mount was prepared from each area (104-52, 134F-52) for which only the rounding index was determined.

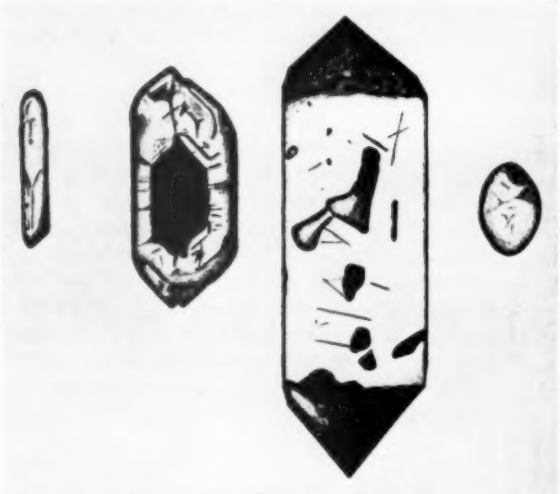
Of the six samples from the Cranberry, four gave elongation-frequency maxima typical of granitic rocks (1F-52, 14-52, 19F-52, 22-52), i.e., maxima above 2.0 (figs. 6 and 7, also table 2). Samples 22-52 and 14-52 have length-frequency maxima typical of igneous granites as well as elongation maxima well above 2.0. In both cases the rounding index is low and clearly favors an igneous origin. The field relations of the two rocks also agree with this interpretation. Sample 22-52 was collected from the southeast side of Doe Hill and is a coarse-grained, porphyritic, biotite granite gneiss. The rock is coarsely foliated but shows no signs of banding. Sample 14-52, collected from the southwest side of the Crossnore quarry, 0.7 mile northwest of Crossnore on the road to Mt. Pleasant Church, is also a coarse-grained, porphyritic granite gneiss. Plate 4 contains a drawing of a zircon from 14-52 showing the typical sharp terminations found in this sample and 22-52. In each case the contact of the foliated granitic rock with the surrounding metasediments is covered.

The elongation-frequency maxima of 1F-52 and 19F-52 are 2.4 and 2.2 respectively, while the length-frequency maxima are 0.16 mm and 0.13 mm. The 0.13 mm value is in the range of zircons of arenaceous rocks while the 0.16 mm maximum is just above this range (Poldervaart, 1955). The rounding index of 19F-52 is 94 percent. In this specimen, rounding of individual crystals is slight, thus preserving the original prismatic habit of the zircons. Sample

TABLE 2

Sample	Formation	Field description of rock	Maximum length-frequency	Maximum elongation-frequency	Rounding index in %	Genesis of rock on field occurrence	Genesis of rock on zircon data
1F-52	Cranberry	Banded granitic gneiss	.16 mm	2.4	61.5	S	S
11F-52	Cranberry	Mica schist	.10 mm	1.2	97.0	S	S
14-52	Cranberry	Foliated granitic gneiss	.40 mm	2.9	18.0	?	I
19F-52	Cranberry	Banded granitic gneiss	.13 mm	2.2, 3.0	94.0	S	S
22-52	Cranberry	Foliated granitic gneiss	.37 mm	2.5	8.0	?	I
104-52	Cranberry	Foliated granitic gneiss	—	—	8.0	?	I
33F-52	Henderson	Banded granitic gneiss	.07 mm	1.4	98.0	S	S
55F-52	Henderson	Banded granitic gneiss	.12 mm	3.0	92.0	S	S
123F-52	Henderson	Foliated granitic gneiss	.12 mm	2.8	94.0	S	S
134F-52	Henderson	Banded granitic gneiss	—	—	94.0	S	S

PLATE 4



Drawings of typical zircons found in four of the ten samples studied.
A. 55F-52: L = 0.14 mm; W = 0.025 mm.
B. 1F-52: L = 0.22 mm; W = 0.099 mm.
C. 14-52: L = 0.523 mm; W = 0.135 mm.
D. 33F-52: L = 0.085 mm; W = 0.059 mm.

1F-52 (pl. 4) contains both clear and dark brown zircons. In counting only the clear zircons in 1F-52, 61.5 percent show rounding of the same type as observed in 19F-52, and 38.5 percent were euhedral. It appears from the data that both specimens are metamorphosed sediments probably of local derivation, as the elongation-frequency maxima are above 2. Sample 19F-52 was collected from a 200-foot road cut 0.2 mile west of Newland, on the road to Minneapolis. This exposure forms the best example of coarse- and fine-banded Cranberry known to the authors. Sample 1F-52 was collected 0.4 mile north of Crossnore quarry on the west side of the road. The rock is a light-colored, coarse-grained, porphyroblastic granitic gneiss. Coarse banding is prominent with slight foliation. The malacons in the rock (1F-52) are of particular interest. When only malacons are counted a rounding index of 70 percent is obtained, or 8.5 percent more rounded grains than for the clear zircons. In view of this rather small increase in the percent of rounded grains and the presence of very dark to black zircons with perfect euhedral outline, it is not at all certain that color throughout necessarily reflects the degree of metamictization of zircon. According to Poldervaart (1955), malacons are rare in arenites as they are unable to resist the effects of abrasion during transport over normal distances.

Specimen 11F-52 (fig. 6) has an elongation-frequency maximum of 1.2, while the length-frequency falls in the upper part of the siltstone range (0.0-.1 mm). The specimen is a chlorite-biotite schist collected 0.3 mile north of

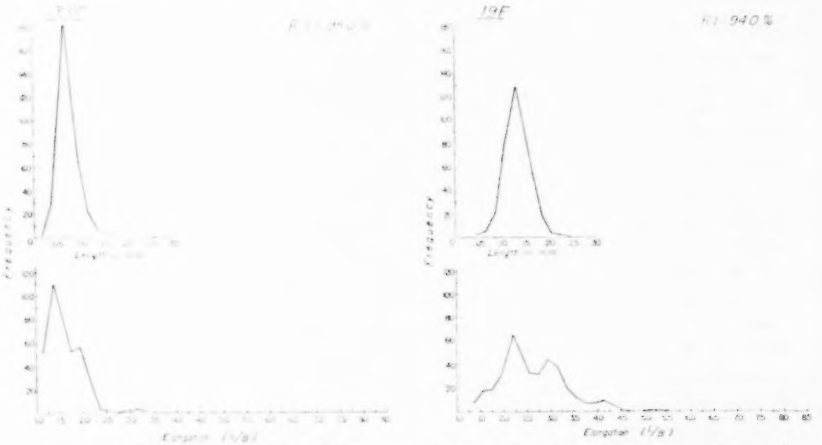


Fig. 7. Elongation- and length-frequency diagrams for zircon concentrates from rock samples 19F-52 and 33F-52.

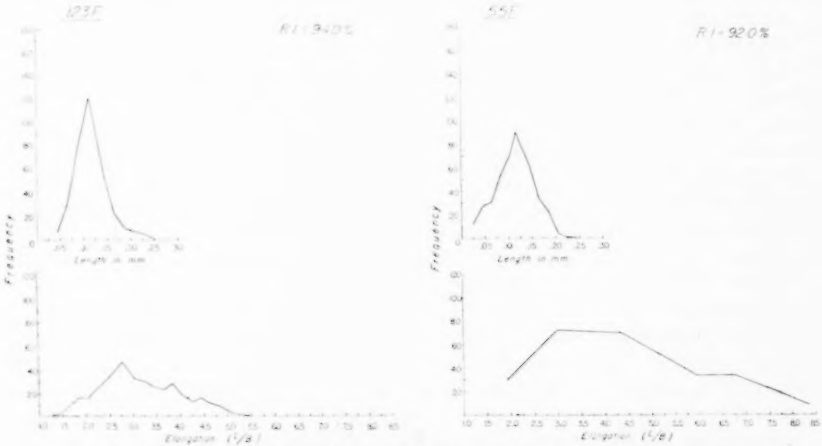


Fig. 8. Elongation- and length-frequency diagrams for zircon concentrates from rock samples 55F-52 and 123F-52.

Crossnore quarry. The rock forms lenses and bands interbedded with quartzofeldspathic gneisses represented by 1F-52. The foliation of the schist is invariably parallel to the banding and foliation of the granitic gneisses.

The final Cranberry specimen (104-52) is a light gray, medium- to coarse-grained porphyritic rock without banding, collected just north of Relief on the road to Pigeon Roost. Of one hundred zircons counted in this sample, 92 percent were euhedral and 8 percent rounded. No other data were obtained from the specimen. The locality may be considered to represent

another local foliated granite mass similar to the Crossnore quarry mass and the east end of Doe Hill. Each is small in extent and surrounded by well-banded granitic gneisses of sedimentary origin.

Hence three of the six specimens collected from the Cranberry formations are clearly of igneous origin on laboratory data. Two specimens represent locally derived arenaceous sediments, while the last sample considered is an argillaceous intercalation in the arenaceous metasediments.

Of the four samples from the Henderson, only one (33F-52), gave an elongation-frequency maximum typical of a sediment. The length-frequency maximum was 0.07 mm while 98 percent of the zircon grains were rounded (pl. 4) and only 2 percent euhedral (fig. 7). Thus the laboratory data for this specimen clearly points to a sedimentary origin. Field observations also agree with this conclusion. The specimen was collected 0.2 mile west on route 26 from the junction of 26 and 26A. The outcrop (approximately 100 feet in length) consists of light-colored (medium- to fine-grained, porphyroblastic gneiss, composed mainly of quartz and feldspar with small amounts of muscovite and biotite. Banding on a coarse and fine scale is prominent while foliation is almost lacking because of the small amount of mica present. Two of the samples (55F-52, 123F-52, fig. 8) present problems similar to those discussed in connection with samples 19F-52 and 1F-52. Sample 55F-52 has an elongation-frequency maximum of 3.0, length-frequency maximum of 0.12 mm, and a rounding index of 92 percent. The majority of the grains show slight rounding at the terminations. Thus despite the high elongation-frequency maximum, the high rounding index clearly indicates that the rock is a psammitic metasediment. Of interest is the wide range over which the elongation-frequency curve is spread, particularly in view of the pronounced 0.12 mm length-frequency maximum (fig. 8). Variations in the width of these extremely thin, pencil-shaped crystals (pl. 4) give rise to the large spread of the elongation-frequency curve. In outcrop the rock is a light-colored, medium- to fine-grained, equigranular gneiss with feldspar and quartz as dominant minerals. Delicate banding on a coarse and fine scale is prominent. DE-123F has a elongation-maximum of 2.8, but a length maximum of 0.16 mm. Of 100 grains counted, 94 percent were rounded and 6 percent euhedral, rounding in most grains being slight, yet clearly visible under high magnification. In outcrop the rock is a light gray medium- to coarse-grained, porphyroblastic gneiss of granitic composition. Foliation is prominent, but banding only faint and discontinuous. Field observations yield few clues as to the origin of the rock. However, the zircon data strongly suggest that the gneiss was originally an arenaceous sediment.

For the last specimen (134F-52) of Henderson gneiss from which zircons were concentrated, only the rounding index was determined. The specimen locality, 2 miles southwest of Old Fort along route 70, has already been cited as affording the best example of banded Henderson known to the authors. Zircons obtained from this specimen were all concentrated in the fine fraction and of 100 grains counted 94 percent were rounded and 6 percent euhedral. The degree of rounding was again slight. The rounding index once again indicates a sedimentary origin.

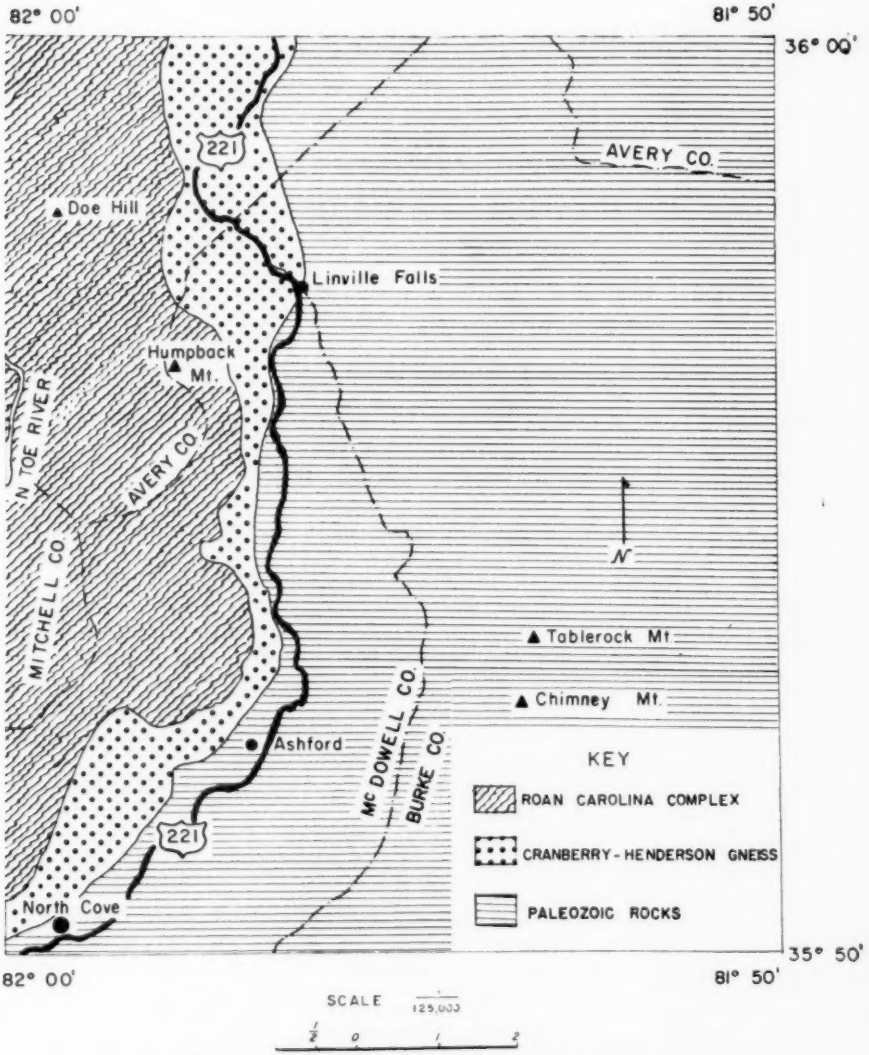


Fig. 9. Detailed map of North Cove-Linville Falls area.

Summary

Zircon concentrates from ten samples representative of the main rock types in the Cranberry-Henderson formation were studied. Of the six samples from the area mapped by Keith as Cranberry, laboratory data favor a sedimentary origin for specimens 1F-52, 11F-52, and 19F-52 and an igneous origin

for specimens 14-52, 22-52 and 104-52. Similar data point to a sedimentary origin for all specimens of Henderson gneiss studied. Thus conclusions as to the genesis of each rock sample based on laboratory data coincide with the conclusions based on field observations alone. In the three cases where the origin was not clearly indicated by field data, the rocks proved to be of igneous origin.

In this study it is evident that the examination of zircon concentrates from granitic gneisses has provided important information bearing on the genesis of the rocks in question. The number of samples considered here is small compared to the area considered and hence they serve only as a means of testing a hypothesis based largely on other observations.

DISCUSSION AND SUMMARY

Sedimentary Origin of the Cranberry and Henderson Gneisses

It is concluded that the sedimentary origin of the Cranberry and Henderson gneisses immediately underlying the Spruce Pine synclinorium is established. The rocks represent the lower part of a thick sedimentary sequence which was subjected to regional metamorphism and varying degrees of feldspathization. The reasons for accepting this mode of origin may be reviewed as follows:

1. Primary compositional banding throughout 80 percent of these formations.

2. The interlayering of Cranberry and Henderson with the Roan-Carolina complex of mica and hornblende gneisses and schists which has been demonstrated to be sedimentary in origin (Brobst and Kulp, in press).

3. Crosscutting relationships between the Cranberry or Henderson and the Roan-Carolina metasedimentary sequence are entirely absent.

4. Conversely the Cranberry and Henderson formations are in a consistent and conformable stratigraphic position below the Carolina-Roan complex as seen in the Cranberry, Roan Mountain and Mount Mitchell quadrangles (fig. 3) where the granitic gneisses are always found to be distributed about the borders of the Roan-Carolina rocks which form the central portion of the southwest-plunging Spruce Pine synclinorium. There is no known locality where the Cranberry or Henderson gneiss outcrops within the main body of Roan-Carolina rocks.

5. Analysis of the size, shape and rounding index of the zircons extracted from these rocks strongly supports the hypothesis that the rocks which display primary compositional banding are sedimentary in origin.

Igneous Phases of the Cranberry and Henderson Gneisses

Crossnore quarry, the east end of Doe Hill, and a road cut immediately north of Relief (on the road to Pigeon Roost) are the only localities known to the authors within the Cranberry-Henderson formation, immediately peripheral to the Roan-Carolina complex of the Spruce Pine area, where intrusive granites outcrop. Provided one considers the intrusion of small granite masses to occur at the end of a period of regional metamorphism, only two occasions exist in the geologic history of this area when one would expect granite intrusion. These times would correspond with the two periods of regional

metamorphism ascribed to the area by Kulp and Poldervaart (in press).

Of the minor constituents present in these rocks, tourmaline is particularly important as it sheds light on the time of probable emplacement. About each tourmaline crystal is a reaction rim of clinozoisite formed at the expense of the tourmaline. Such a phenomenon requires a reheating of the surrounding rock and recrystallization, such as one would expect to happen during regional metamorphism. As this evidence requires the granite itself to have been metamorphosed, its emplacement most probably took place toward the end of the first period of metamorphism (Poldervaart, oral communication) thus allowing the clinozoisite rims about tourmaline crystals to be produced by the second period of regional metamorphism. The granite itself may be a product of the first metamorphism, having originated at much greater depth than its present setting.

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USE OF STATISTICAL METHODS TO DETECT RADIOACTIVITY CHANGE DUE TO WEATHERING OF A GRANITE*

W. L. SMITH and F. J. FLANAGAN

ABSTRACT. Forty-four samples of the Conway granite were collected from the red and green phases of the rock at the Redstone, New Hampshire quarries. A large variation in radioactivity as measured by β-counting is shown between individual samples.

Inspection of the data shows that the red phase is higher in radioactivity than the green. An analysis of variance with a single variable of classification shows that the means of the fresh and weathered red phases are not significantly different, whereas a "t" test using differences between pairs of the fresh and weathered green samples shows that the means of these two sets differ significantly. From these tests and a comparison of the variances of the respective sets, it is inferred that weathering has had a significant effect on the green phase only.

It has been shown, by comparing the variances of the subsets of data with the known variance of the method of measurement, that some external factor such as variations in mineralogic composition or differential leaching or adsorption may be responsible for the variations in radioactivity.

INTRODUCTION

A study of the radioactivity and the distribution of uranium in the rocks and minerals of the Conway granite of New Hampshire has been undertaken under the direction of A. P. Butler, Jr. of the U. S. Geological Survey. Replicate uranium determinations of an analytical sample split from a previously obtained 40-pound sample of this granite gave results whose variance greatly exceeds (by a factor of about 10) the known variance of the analytical determination. As this large variance was obtained from replicate determinations of the same sample, the possibility exists that this variance may be equal to or greater than any difference in the uranium content of the granite that we might wish to detect. It is therefore necessary to determine by some other means whether differences in the granite exist.

Radioactivity determinations by β-counting should be relatively unaffected by sampling errors as the size of the sample for analysis is about 100 grams compared to a 1-gram sample for chemical analysis. It was believed therefore that small differences in radioactivity might be detectable and might be used as a guide for other studies.

It was further believed that, as a part of the general study of the Conway granite, some measure of the variation of radioactivity for closely spaced samples would be useful for comparison as more data became available. For this study, William Lee Smith, Rachel M. Barker, and Regina L. Wack of the Geological Survey collected a number of 5-pound samples of the granite at the Redstone, Carroll County, New Hampshire quarries. The samples were collected for the immediate purposes of determining (1) the variation of radioactivity in the two phases of the Conway granite, and (2) the effect of weathering on the radioactivity of the rocks.

* Publication authorized by the Director, U. S. Geological Survey.

ACKNOWLEDGMENTS

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THE CONWAY GRANITE

The Conway granite of Mississippian (?) age is one of a series of the alkalic rocks which comprise the White Mountain batholith. The White Mountain batholith includes both intrusive and extrusive phases. At Redstone, New Hampshire a red phase and a green phase of the Conway granite have been mined by the Maine and New Hampshire Granite Corporation from adjacent quarries since 1889. The two phases are separated by a fault contact which is at the west end of the Redstone Red quarry. The distribution of the Conway granite in this area has been described by Billings (1928), and the petrography has been described by Billings (1928) and by Dale (1908).

The Conway granite is relatively high in uranium content for its rock type and contains proportionately large amounts of accessory minerals. The red phase of the granite is a coarse even-grained biotite granite containing microperthite, smoky quartz, biotite, amphiboles, and accessory minerals as shown in table 1.

The green phase of the Conway granite is a coarse-grained yellow-green granite composed mainly of green microperthite, smoky quartz, amphiboles, and biotite. Billings (1928) describes the green phase as a segregation from the red phase. Comparison of the mineral compositions in table 1 shows the difference between the fresh red and fresh green rock.

TABLE 1

Minerals of Conway Granite at Redstone, New Hampshire

Mineral	Red phase	Green phase
	Major components	
Feldspar	70	75
Quartz	25	20
Biotite	3.0	0.2
Amphiboles	0.2	3.0
	Accessory minerals (in order of abundance)	
	Magnetite	
	Fluorite	Fluorite
	Allanite	Allanite
	Zircon	Zircon
	Apatite	
	Ilmenite	Pyrite
	Pyrite	Thorite

Analyses of the two phases by W. H. Herdsman (Billings, 1928) are shown in table 2.

TABLE 2
Chemical Analyses

	Red phase	Green phase
SiO_2	70.45	69.85
Al_2O_3	14.87	14.44
Fe_2O_3	1.07	0.94
FeO	1.54	2.37
MgO	0.35	0.24
CaO	1.28	1.25
Na_2O	4.28	3.93
K_2O	4.16	5.04
H_2O+	0.65	0.66
H_2O-	0.50	0.35
TiO_2	0.69	0.74
P_2O_5	0.04	0.03
MnO	0.00	tr
Total	99.88	99.84
Sp. gr.	2.63	2.65

Weathering has altered the Conway granite near the glaciated surface of Redstone Ridge, producing a greater amount of iron oxide staining toward the top of the quarries. Billings (1928) describes completely disintegrated glacial boulders of the red phase 1 foot in diameter and concludes that the breakdown of the rock has occurred since Wisconsin time. Dale (1908) describes limonite, kaolinite, calcite, and epidote as secondary minerals in the green phase; and hematite, kaolinite, leucoxene, chlorite, and calcite as secondary minerals in the red phase.

COLLECTING METHOD

Forty-four samples of the Conway granite were collected from both quarries. In the Redstone Green quarry a rock face 200 feet long by 50 feet deep is exposed; the Redstone Red quarry exposes 500 feet of granite to depths ranging from 30 to 80 feet. The samples were collected at regular intervals along the quarry faces in both the fresh rock at lower levels and in the weathered zone. The samples of weathered rock were collected approximately vertically above the fresh samples.

METHOD OF MEASUREMENT

All samples were ground to pass an 80-mesh screen, and their radioactivity was measured by a routine β-γ method. The apparatus consists of a glass tube-and-sample holder made of a complete ground-glass joint into the bottom half of which is sealed securely in a coaxial position a 30 mg/cm^2 thin-walled Geiger tube. The outer half of the joint, when secured in proper position on the inner half, surrounds the Geiger tube so that the powdered samples may be held in a reproducible position. The unshielded tube-and-sample holder is held vertically by a ring stand and clamp and is connected to a conventional scaler. Background determinations are made using a radioactively inert salt

such as sodium sulfate, and standardization is accomplished by comparison with counted chemically analyzed standards.

The usual assumptions are made that (1) all the radioactivity of the samples is due to uranium and its daughter products, and (2) the uranium is in secular equilibrium with its daughter products. Although these two assumptions may not be true, they allow calculations of radioactivity measurements in terms of the percent equivalent uranium (eU), that is, the radioactivity which is equivalent to that shown by a specified percentage of uranium in equilibrium with its daughter products. This method of calculation has the advantage that intra-laboratory comparisons may be made without the laboratories specifying their exact counting conditions.

DATA AND CALCULATIONS

Radioactivity determinations were made on the samples, and the results, expressed as net β-γ counts per 5-minute counting period, are shown in table 3.

Preliminary inspection of the data shows that: (1) a radioactivity count, for sample 136, of 1418 is about twice the value of the mean of all others in its class, and (2) the weathered green phase is higher in radioactivity, with one exception, than the fresh green, whereas the high and low values for the red phase are about equally distributed between the weathered and the fresh rocks. Although the value of 1418 counts above may be rejected upon inspection alone, it may be strongly rejected from a calculated "t" value of approximately 33, calculated by using the mean and the standard deviation of the other nine samples in the weathered green phase group. This value of t is obtained by substituting the mean $\bar{x}$ and its standard deviation s, together with the questioned observation x in the equation $t = \dfrac{x - \bar{x}}{s/\sqrt{n}}$ where n is the number of observations. The table "t" value for the 95 percent confidence level for eight degrees of freedom is 2.31.

It seems from the pattern of sample collecting at regular intervals that the data could be treated by an analysis of variance in a 2×2 classification design where color would be one classification and the condition of the rock the other. One of the prerequisites for the analysis of variance is that the variances of the subsets be homogeneous, and this homogeneity, or the lack of it, may be shown by Bartlett's test for homogeneity of variance given in any elementary statistics book. Pearson and Hartley (1954) have calculated a table to reduce the computations necessary for Bartlett's test. These tables show values for the maximum allowable ratio of the maximum to the minimum variance of k sets of variances where each variance has v degrees of freedom. The ratio for our four sets of data, $12260/507 = 24.2$, exceeds the tabled value of 5.67. The variances are therefore heterogeneous and an analysis of variance using the four sets of data may not be used.

We may apply the same test to the variances of the weathered and fresh rocks of either phase and we find that only the red phase has variances that are homogeneous. Having assured ourselves that the data for the red phase are normally distributed, we may then use the analysis of variance with a

TABLE 3

Radioactivity of Redstone Quarry Samples

Green phase				Red phase			
Weathered		Fresh		Weathered		Fresh	
Sample	Counts*	Sample	Counts*	Sample	Counts*	Sample	Counts*
135	907	113	710	123	836	101	942
136	(1418)**	114	754	124	768	102	887
137	799	115	712	125	904	103	1041
138	831	116	750	126	1044	104	933
139	742	117	713	127	1143	105	969
140	716	118	767	128	953	106	1023
141	801	119	715	129	1030	107	1126
142	760	120	700	130	924	108	776
143	806	121	724	131	996	109	980
144	821	122	716	132	893	110	1051
				133	1100	111	1081
				134	873	112	1060
Number		9	10	12		12	
Mean		798.1	726.1	955.3		989.1	
Variance		3109.	507.	12260.		9234.	
Standard deviation		55.8	22.5	110.7		96.1	

* Net β-γ counts in five minutes.

** This value omitted from calculations. (See discussion above.)

single variable of classification, calculations for which are shown in Dixon and Massey (1951). The analysis of variance table given below shows that there is not a significant difference between the means of the weathered and fresh red phases.

	Sum of squares	d.f.	Mean square	F ratio
Means	6834	1	6834	$F = 6834/10747 = 0.64$
Within	236436	22	10747	$F_{.05}(1,22) = 4.30$
Total	243270	23		

As the variances of the weathered and fresh green phases are heterogeneous and cannot be treated by an analysis of variance, we may treat these data by pairing the observations. This technique, useful in agriculture where a series of pairs of plots have one plot of the pair treated with fertilizer and the other untreated, has the advantage that any differences from pair to pair have probably been eliminated by taking differences between pairs, and we need not assume that the variances of the two populations under the test are equal.

The hypothesis we wish to test is that the means of the weathered and fresh green samples are equal. To test this hypothesis we may use Student's "t" test in the form

$$t = \frac{\bar{d}}{s_d/\sqrt{n}}$$

where $\bar{d}$ is the average difference between pairs and hence equal to the difference between the means of the two groups, s_d is the standard deviation of the

differences and n is the number of pairs available. We will choose the 95 percent confidence level for 8 degrees of freedom for which t = 2.31. Substitution of the calculated quantities in the equation yields a t equal to 3.44. We therefore reject the hypothesis at the 95 percent level that the means of the fresh and weathered green samples are equal. The hypothesis would also be rejected at the 99 percent level, or the chance that the two means are equal is less than 1 in 100.

We may also pair the counts for the fresh and weathered red sets, take the differences between pairs, and apply the "t" test in a similar manner. The calculated result, t = −1.80, does not exceed the table value for a one-sided test, $t_{0.05}$(d.f. 11) = −2.20, and hence the difference between the means of the two groups is not significant.

We could compare the four subsets of data by testing certain hypotheses concerning pairs of variances. If the variances of two sets of random samples are not significantly different, then it may be inferred that the variances of the populations from which the two sets of random samples were drawn are not significantly different. The statistic used for this test is the F ratio, $F = \frac{s_1^2}{s_2^2}$, where s_1^2 is the larger of the two variances. The hypothesis we wish to test is that the variances σ^2 of the different populations, from which samples with variances s^2 were drawn, are equal. The 95 percent confidence level, i.e., one chance out of twenty being wrong, has been chosen. The sample pairs whose variances are being tested, the F ratios at the 95 percent confidence level, and the calculated F ratios are shown in table 4.

There is therefore only one set of variances that may be considered as having come from the same populations and that set is the weathered red-fresh red pair. All other combinations have variances significantly different; the fresh red-fresh green pair are very significant.

Consideration of the data in table 3 shows that a considerable variation exists in the radioactivity of the red-phase samples as compared to the green.

TABLE 4
Comparison of Variances

Variance ratio s_1^2/s_2^2	$F_{.05}(N_1-1,N_2-1)$	F (calculated)	Conclusion
Weathered green/fresh green	3.23	$\frac{3109}{507} = 6.13$	Significant
Weathered red/fresh red	2.82	$\frac{12260}{9234} = 1.33$	Not significant
Weathered red/weathered green	3.32	$\frac{12260}{3109} = 3.94$	Significant
Fresh red/fresh green	3.10	$\frac{9234}{507} = 18.21$	Significant

The variation in radioactivity in the red phase, and to a much lesser extent in the green, leads one to suspect that one or more unknown factors are in-

fluencing the radioactivity. This effect might be due to several causes, for example, different mineralogic composition, differential leaching, or absorption of radioelements.

This hypothesis of possible causes for the variation is further strengthened by data obtained in the laboratory. Radioactivity measurements made during an internal standardization of a shale sample with the same counting intervals showed that this shale sample had a mean radioactivity of 0.0058 percent eU with a standard deviation of $\pm$ 0.00016 percent eU or a coefficient of variation of $(\frac{0.00016}{0.0058} \times 100)$ 2.8 percent. Uranium is believed to be the main source of radioactivity in this shale and it is generally believed to be uniformly distributed horizontally. The standard deviation above obtained from 100 randomly selected portions of the same sample, counted on two different counters and by two different operators, may then be called an estimate of the population standard deviation. As the 100 portions were merely replicate determinations of the same shale sample, we may also call this estimate the expected standard deviation, σ_m, of the method where the counting rate is about 1000 counts per five-minute counting period.

This now gives a basis for comparison. If the samples of the red phase are homogeneous with respect to the source of radioactivity, a standard deviation of 28 counts would be expected. This is obtained by multiplying the mean of the counts of the red phase, 989, by the known coefficient of variation, 2.8 percent. We shall assume that the variance, σ^2, shown by the fresh red phase samples, for which s^2 of the samples is an estimate, was obtained from a random sample of the population from which our method variance, σ^2_m, was derived. To test this hypothesis we shall use the statistic $\chi^2/\text{d.f.} = \frac{s^2}{\sigma_m^2}$ which has a $\chi^2/\text{d.f.}(N\text{-}1)$ distribution where the degrees of freedom (d.f.) are equal to (N-1), N being the number of observations in the sample. The upper value of $\chi^2/\text{d.f.}$ (11) at the 95 percent confidence interval is 1.79. The calculated $\chi^2/\text{d.f.}$ (11) is equal to $\frac{9234}{(28)^2}$ or 11.78, which is extremely significant.

As the same method of measurement was used for both the shale and the granite, the observed differences must be due, not to the method, but to variations in the samples themselves. Similar calculations involving the observed variances of the other three groups of samples show that only the variance for fresh green phase samples may be considered as being the same as the methods variance, σ_m^2. It must be pointed out that, whereas the significant values of $\chi^2/\text{d.f.}$ may indicate a possible difference in mineralogic composition, the nonsignificance of $\chi^2/\text{d.f.}$ for the fresh green phase does not indicate that small differences in the samples may not exist.

The sample data with radioactivity values calculated to percent equivalent uranium (eU) are given in table 5.

TABLE 5

Radioactivity of the Conway Granite, Redstone Quarries, New Hampshire

Green phase				Red phase			
Weathered		Fresh		Weathered		Fresh	
Sample	eUx10^4	Sample	eUx10^4	Sample	eUx10^4	Sample	eUx10^4
135	52	113	41	123	48	101	55
136	82	114	44	124	45	102	51
137	46	115	41	125	52	103	60
138	48	116	46	126	61	104	54
139	43	117	41	127	66	105	56
140	41	118	44	128	55	106	59
141	47	119	41	129	60	107	65
142	44	120	40	130	54	108	45
143	47	121	42	131	58	109	57
144	48	122	41	132	52	110	61
				133	64	111	63
				134	51	112	61

DISCUSSION AND SUMMARY

One of the more important reasons for the close sampling of the Conway granite in the Redstone quarries was to determine if small differences in radioactivity could be detected, and such differences between the groups of samples have been noted.

It is important to note that although we discarded the observation of 1418 counts for sample 136 for statistical reasons, it did occur in one of ten samples of the selected lot. Sufficient data are not available to predict the probability of picking such a sample from an infinite number, but obviously caution is indicated in claiming significant differences based on single samples over large areas.

It has been determined, using the analysis of variance with a single variable of classification, that there is not a significant difference between the means of the weathered and fresh rock of the red phase, and it has been shown that the variances of the two sets of samples are not significantly different. Although the ratio is compared at the 95 percent level, reference to tables of F ratios at various confidence levels shows that, for the calculated ratio 1.33, the variances are not significantly different at the 70 percent level. This gives added assurance to the inference that the populations from which these sets of samples were taken might be the same.

The hypothesis that the means of the fresh and weathered green samples are equal, or that the differences between the two means is zero, was tested using the average difference and its standard deviation in a "t" test. The conclusion reached is that the average difference is significantly different from zero and hence the means are not equal. It was then inferred that the populations from which the samples were drawn are significantly different and this inference is supported by a comparison of the calculated variance ratio to the table value.

The red phase shows variances significantly different from those of the green regardless of the rock condition, and from this we may infer that the samples, and hence the populations, of the two phases are significantly different.

Having drawn inferences about the different populations after comparing means or variances, it has been shown by comparing the variances of the subsets of data with the known variance of the method of measurement that only the variance of the fresh green rock is not significantly different from that of the measurement. As a result of this, the logical inference to be made is that external factors such as different mineralogic composition or differential leaching or adsorption are responsible for the variances significantly different from that of the method.

From a geologic viewpoint, the most important conclusion that can be drawn from the data and based on the statistical inferences made is that weathering has affected the green phase of the Conway granite significantly whereas any effect on the red phase is not apparent.

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REVIEWS

Astrophysics; by LAWRENCE H. ALLER. *Vol. 1, The Atmospheres of the Sun and Stars.* P. ix, 412; 118 figs. *Vol. 2 Nuclear Transformations, Stellar Interiors, and Nebulae.* P. x, 291; 51 figs. New York, 1953 (Ronald Press Company, $12.00 each volume).—This book fills a gap in the astrophysical literature. Although there is no lack of monographs appealing to specialists and advanced students, this is the first textbook in the English language that is designed for general use by graduate students and treats all branches of astrophysics. Considering the wide range of topics covered, Dr. Aller has achieved a remarkable balance. Unusual in a textbook are the copious references to the periodical literature, assembled at the end of each chapter. Immense care has evidently been given to the preparation and classification of these references, for they are distinctly complete and up to date. Even research workers will find them useful for purposes of general orientation outside their special fields. Each chapter is followed by a list of problems. Their quality is rather uneven. Quite a few of these problems, in fact, are merely exercises in algebra, which do not afford a real test of the student's grasp of the underlying physical theories.

Volume 1 begins with a review of basic astronomical data and relevant parts of physics, which takes up about one-third of the entire space. Chapters 2 to 5 deal, respectively, with atomic and molecular spectra, gas laws and the equation of state, ionization and dissociation, emission and absorption of radiation. Certain elementary physical theories, perhaps, could have been recapitulated less broadly. Instead more space could profitably be allotted to more difficult physical concepts, some of which are introduced in a rather off-hand manner. After a brief summary (chap. 6) of the observational characteristics of stellar radiation, chapter 7 brings the theory of continuous stellar spectra and model atmospheres. Chapter 8, on the Fraunhofer spectrum, is the longest one (100 pages). Although in chapters 7 and 8 liberal use is made of the mathematical theory of radiative transfer, students will welcome the clear distinction between mathematics as a tool and the physical core of the astrophysical theories. The final chapter, on solar phenomena (sun-spots, prominences, chromosphere and corona), contains many excellent half-tone illustrations. There is also a brief description of radio-frequency radiation from the sun and of solar-terrestrial relationships.

Volume 2 opens with a discussion of the physics of stellar interiors (radiative vs. convective equilibrium, stellar absorption coefficient and energy generation) and then proceeds to the computation of stellar models (chap. 2), which is illustrated by a fairly detailed numerical example. There are brief sections on stellar evolution and the origin of the chemical elements. Chapter 3 deals with Cepheids and long-period variables and their explanation in terms of the pulsation theory. The summary of relevant observational data might have been enlivened by reproducing some of the instructive diagrams, due to Milne and his collaborators, which show the closed paths traced by these variables on the Russell diagram. These beautiful graphs seem to have fallen into oblivion but deserve to be more widely known. Chapter 4 takes up several types of unstable stars (novae, supernovae, SS Cygni-variables), whose physical interpretation is still beset with difficulties. The nature of planetary

nebulae, to the study of which the author has made notable contributions, is analyzed in chapter 5. The concluding chapter on the interstellar medium (including galactic radio-frequency radiation) seems disproportionally short.

For an introductory text the high price of these two volumes is disconcerting. But paper and typography are excellent. It is to be hoped that Dr. Aller will before long enjoy the satisfaction of having to prepare another edition. RUPERT WILDT

Petrographic Mineralogy; by ERNEST E. WAHLSTROM. P. vii, 408; 178 figs. New York, 1955 (John Wiley & Sons, $7.75).—Professor Wahlstrom explains in a brief preface that he wrote this text to fill the gap between works dealing with the theory and principles of the polarizing microscope and those dealing with the classification and properties of minerals and rocks. In this single volume he has attempted to bring together all the data necessary for the correct identification and classification of the important minerals and rock types as examined under the microscope and by various other petrographic methods. It is designed for a course in optical petrography of the type that would normally follow a course in the theory and operation of the polarizing microscope.

The first five chapters of the book take up a multitude of petrographic techniques and tell the student how to "Do it yourself"; among many other things, this includes staining techniques, heavy liquid separations, modal analyses, operation of the universal stage, CIPW norm calculation, and even (for the student bold enough to attempt it) an outline of thin-section grinding technique. The 24-page section relating to the universal stage is worthy of special mention, for it provides an admirable introduction to a technique which can be especially confusing for the novice petrographer.

Chapters 6 and 7 provide well organized description of silicate and nonsilicate rock-making minerals, with numerous accompanying tables, sketches, large size photomicrographs (3 x 4 inches). Characteristics of each mineral under plane polarized light, crossed nicol prisms, and in hand specimen are given under separate headings, and each mineral description leads off with complete optical data in handy tabular form. Chapter 8 provides 46 pages of determinative tables for approximately 200 rock-forming minerals. These tables are well suited for convenient use in the petrographic laboratory, and in their main outlines are like a simplified and abridged version of the tables to be found in Tröger's unsurpassed "Tabellen zur optischen Bestimmung der gesteinbildenden Minerale."

The final chapters deal with the composition, properties, and classification of igneous, sedimentary, and metamorphic rocks. The emphasis is placed upon description and classification based upon thin-section study, so perhaps it is unfair to criticize the author for recommending a scheme for megascopic classification that lumps all quartz-bearing phaneritic rocks with dominant feldspar together as "granite" and feldspathic rocks without quartz as "syenite." The classification scheme recommended for thin-section study is adequate, but the outline form in which it is presented seems a bit cluttered to the eye; this is only a superficial difficulty, however, for the basic organiza-

tion is sound. The author has wisely cut odd varietal rock names to a minimum, and it is only in his treatment of the lamprophyre group that one receives a full dose of them. Many readers will probably object to the use of "dense" as a grain-size classification for aphanitic rocks on the basis that it more properly refers to high specific gravity than very small grain size.

The value of this book is greatly enhanced for the inquiring student by the rather large number of literature references in the text; the lists of selected references found at the end of each chapter total more than 100 entries. Numerous large size photomicrographs of thin sections contribute a great deal to the handsome appearance of the book, and there are a number of striking pairs of photographs, each pair occupying a full page, showing the same area under plain light and under crossed nicols.

Petrographic Mineralogy makes a fine companion volume to Professor Wahlstrom's earlier text *Optical Crystallography*, and will certainly find wide favor as a text and laboratory "friend" for students in optical petrography courses throughout the country. PHILIP ORVILLE

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and C. A. Hickcox ($ .65). 996-E, Surficial Geology of the Louisville Quadrangle, Colorado; by H. E. Malde ($ .75). 1000-B, Geochemical Prospecting Investigations in the Nyeba Lead-Zinc District, Nigeria; by H. E. Hawkes ($ .25). 1000-C, A Paleozoic Biochemical Anomaly near Jerome, Arizona; by L. C. Huff ($ .15). 1003-A, The "Clinton" Sands in Canton, Dover, Massillon, and Navarre Quadrangles, Ohio; by J. F. Pepper, W. de Witt, Jr., and G. M. Everhart ($1.50). 1008, Geology and Mineral Deposits of the James River-Roanoke River Manganese District, Virginia; by G. H. Espenshade ($6.66). 1009-E, Stratigraphy of the Morrison and Related Formations, Colorado Plateau Region, A Preliminary Report; by L. C. Craig et al. ($ .20). 1009-F, Glossary of Uranium-and-Thorium-Bearing Minerals; by J. W. Frondel and Michael Fleischer ($ .20). 1010, Geologic Controls of Lead and Zinc Deposits in Goodsprings (Yellow Pine) District, Nevada; by C. C. Albritton, Jr., A. Richards, A. L. Brokaw, and J. A. Reinemund ($4.75). 1015-C, Geology of the High Climb Pegmatite Custer County, South Dakota; by D. M. Sheridan ($ .65). 1015-E, Strippable Lignite Deposits Slope and Bowman Counties, North Dakota; by R. C. Kepferle and W. C. Culbertson ($1.00). 1015-F, Geology and Coal Resources of the Henryetta Mining District, Okmulgee County, Oklahoma; by R. I. Dunham and J. V. A. Trumbull ($1.00). 1015-G, Zince-Lead-Copper Resources and General Geology of the Upper Mississippi Valley District; by A. V. Heyl, E. J. Lyons, A. F. Agnew, and C. H. Behre, Jr. ($ .70). 1015-H, Preliminary Geochemical Studies in the Capiton Reef Area, Wayne County, Utah; by L. C. Huff ($ .15). 1019-C, Annotated Bibliography and Index Map of Barite Deposits in United States; by B. G. Dean and D. A. Brobst ($ .45). 1020-A, Geology and Coal Resources of the Connel City Quadrangle, Kentucky; by K. J. Englund ($1.25). 1022-D, Geophysical Abstracts 159, October-December 1954; by M. C. Rabbitt, D. B. Vitaliano, S. T. Vesselowsky et al. ($ .25). 1024-, Radioactivity Investigations in the Cache Creed Area Yentna District, Alaska, 1945; by G. D. Robinson, H. Wedow, Jr., and J. B. Lyons ($ .65). 1027-A, Physical Stratigraphy of the Phosphoria Formation in Part of Southwestern Montana; by E. R. Cressman ($ .55). 1030-A, Search for Uranium in the United States; by V. E. McKelvey ($ .26). 1033-A, Geophysical Abstracts 160, January-March 1955; by M. C. Rabbit, D. B. Vitaliano, S. T. Vesselowsky et al. ($ .25). 17036-A, Rapid Field and Laboratory Method for the Determination of Copper in Soil and Rocks; by H. Almond ($ .10).

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